Bulletin

of

The Geological Society of America

VOLUME 62 June 1951 NUMBER 6

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Subscription $15 per year.

Publication Office: Mt. Royal & Guilford Aves., Baltimore 2, Md.

Communications for publication should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117th Street, New York 27, N. Y.

NOTICE.—Four weeks notice of change of address is necessary. In accordance with the rules established by the Council, claims for non-receipt of the preceding number of the Bulletin must be sent to the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter in the Post-Office at Baltimore, Md., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103, Act of October 3, 1917, authorized on July 8, 1918

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VOL. 62, PP. 547-558, 7 FIGS. JUNE 1951

VARIATIONS IN CHEMICAL COMPOSITION ACROSS IGNEOUS CONTACTS

By WILLIAM H. DENNEN

ABSTRACT

The contact effects resulting from the intrusion of igneous material into various wall rocks are linear compositional changes only in special instances. Large variations in chemical composition from point to point may be present in both intrusive and wall rocks.

A study of these variations by the analysis of closely spaced samples with quantitative spectrographic methods indicates that all of the major rock constituents studied (Al, Fe, Mg, Mn, Ca, Na, and K) are mobile when moderate temperatures and concentration gradients are present. Migration of material is believed to take place late in the cooling history by diffusion along mineral boundaries. The compositional variations resulting from this transfusive migration of hydrated ions are often sufficiently large to mask any compositional changes due to earlier reactions.

The chemical variations found can be related to the original compositions of the wall rock and intrusive material, and to the cooling history of the system. No variations in chemical composition are found across contacts between chemically identical rocks. Generally smooth and continuous variations occur across contacts between rocks of different original composition which have been held at high temperatures. Highly irregular variations are found when the intrusive and wall rock were compositionally different and steep thermal gradients existed. These latter variations may show geochemical culminations and depressions are wave-like in form, and die out in both directions from the contact zone.

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INTRODUCTION

The invasion of igneous materials into various country rock upsets the existing conditions and sets the stage for a complex series of interactions tending toward the establishment of a new equilibrium picture, the resultant changes being primarily controlled by the chemical composition, temperature, and temperature distribution of the system. The compositional changes in the country rock of the contact zone

may vary in magnitude from visually undetectable effects to wide gradational contacts, and from pure recrystallization to complete metasomatic replacement. Similarly, the invading materials undergo changes.

TABLE 1.—REPRODUCIBILITY OF ANALYTICAL METHOD

Oxide	Standard deviation (%)	Number of samples analyzed
Al_2O_3	2.5	152
total iron as Fe_2O_3	4.7	163
MgO	6.5	148
MnO	4.6	151
CaO	5.3	189
Na_2O	6.0	165
K_2O	8.8	72

METHOD OF INVESTIGATION

In the present work, some 20 contacts have been studied by spectrochemical techniques in an attempt to determine the variation present across these critical zones and to obtain some idea about the factors which controlled the variation. The spectrochemical techniques used were those proposed by Kvalheim (1947) as modified by Ahrens (in preparation). The contacts were "traversed" by a suite of specimens taken along a line perpendicular to the observed contact and plots of weight per cent concentration of oxide vs. linear distance from the contact were constructed from the analytical data. The percentage concentration was plotted logarithmically in order to bring out the relative rather than the absolute changes in the oxides. All the contacts analyzed were between initially homogeneous rocks in order to eliminate, so far as possible, any pre-intrusive variations. The curves obtained illustrate the variation of the individual oxides with respect to the contact and the variation of the oxides with respect to each other—that is, their geochemical affinities.

Samples for analysis were prepared from rock slices cut from large specimens showing a contact. The cuts were made parallel to the contact and spaced so that each slice would contain a minimum of 1000 grains. Samples at a distance from a contact were obtained by chip sampling.

The analytical errors inherent in the spectrographic method were determined by computing the standard deviation for each analysis, which was made in triplicate, and constructing a distribution curve of the standard deviations obtained. The standard deviation value of the distribution curve peak was taken as the overall analytical reproducibility for this method (Table 1).

COMPOSITIONAL CHANGES IN A HOMOGENEOUS ROCK

The possible contribution of compositional differences in an apparently homogeneous rock to the variation observed in trans-contact samples was investigated by a series of analyses of a single rock. A porphyritic, foliated, and micaceous granite from Chelmsford, Mass., was utilized. A 1 x 1 inch column, cut perpendicular to the foliation, was sawed into 1-inch cubes and alternate cubes were analyzed. The samples were somewhat undersize and compositional banding is present.

The relatively slight amount of chemical variation from point to point in this rock (Fig. 1) is the algebraic sum of (a) initial inhomogeneity, (b) analytical and sampling errors, and (c) contamination.

The homogeneity of the Chelmsford granite is less than that of the rocks chosen for analysis and this granite should, therefore, show a point to point variation greater than may be expected in more nearly homogeneous rocks. Thus, stronger variations within such rocks than are shown in Figure 1 must be attributed to another cause.

The absolute accuracy of the weight per cent values has not been considered. These values are known to deviate from the true values for certain oxides, for example Al_2O_3 in Figure 4, but this deviation is a constant factor for all determinations and thus does not affect the shape of the variation curves, which are based on relative changes.

CLASSIFICATION OF CONTACTS

The parameters controlling migration of material across hot contacts are temperature and temperature gradient, the chemical composition of the materials involved, pressure, and time.

Of these, a general magnitude for temperature and approximate composition may be inferred from geologic considerations with reasonable certainty at any given contact. The examples studied have consequently been classified as follows:

A. Invading and invaded materials compositionally similar.
   1. High temperature, but no thermal gradient.
   2. Steep thermal gradient.
B. Invading and invaded materials compositionally dissimilar.
   1. High temperature, but no thermal gradient.
   2. Steep thermal gradient.

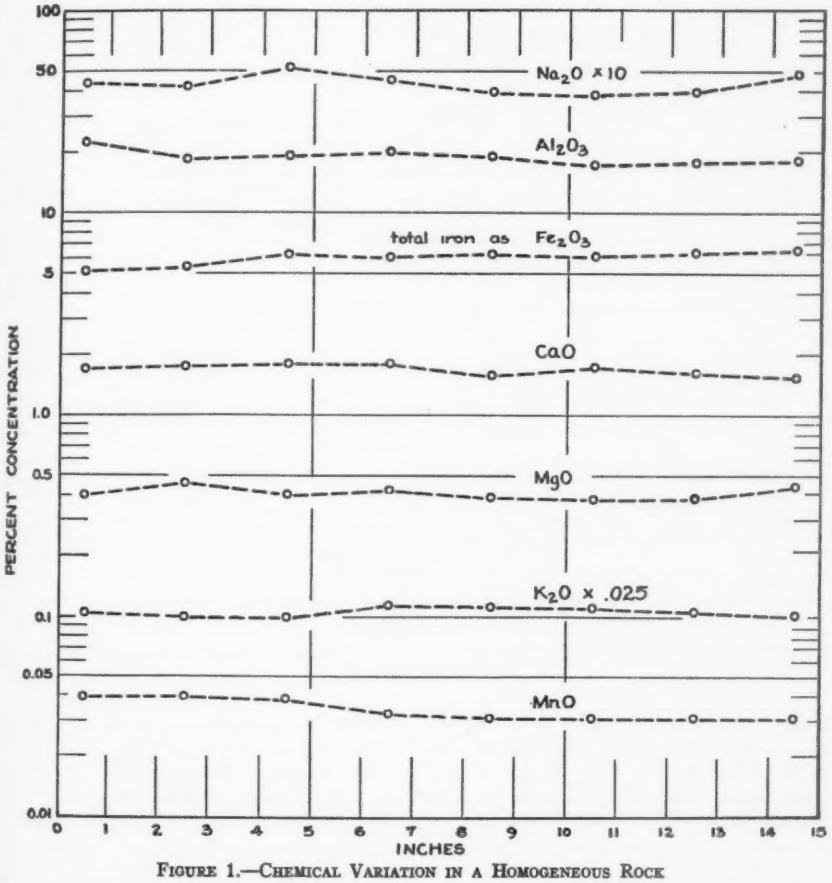

FIGURE 1.—CHEMICAL VARIATION IN A HOMOGENEOUS ROCK

CONTACTS OF COMPOSITIONALLY SIMILAR ROCKS

High temperatures alone are ineffective in the net transfer of materials across a contact during the emplacement of igneous materials unless the invading and invaded rocks are chemically dissimilar. This is difficult of proof since intrusive relations between chemically identical rocks are seldom observed in the field. None of the contacts studied fit the qualifications of compositional similarity and a high temperature for all constituents. When, however, any element is present in equivalent amounts in both invading and invaded phases, small

transcontact variation is usual, and may be seen for soda and alumina in Figure 3.

The condition of equivalent composition and temperature difference was closely approximated in a set of samples taken across the contact of a 52-inch trap dike intrusive into hybridized gabbro-diorite at Little Nahant, Massachusetts. The contact is wavy in detail with apopheses a few inches long extending from the dike. The dike shows a chilled zone about 1 cm. thick. The plot of this contact (Fig. 2) shows slightly more variation in the major elements from point to point than is the case for a homogeneous rock (Fig. 1), but considerably less variation than is found across the contact of rocks of dissimilar composition.

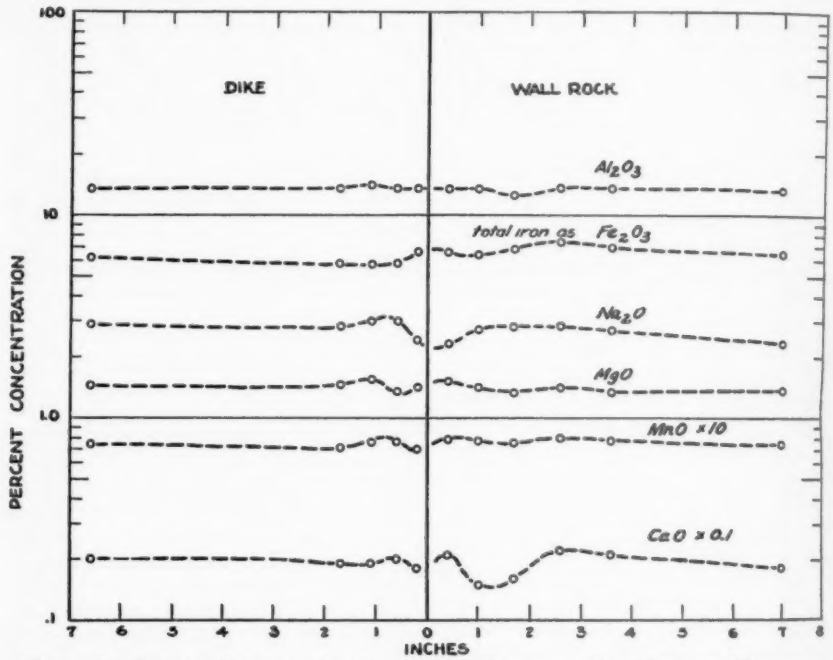

FIGURE 2.—CHEMICAL VARIATION ACROSS A CONTACT BETWEEN A TRAP DIKE AND HYBRIDIZED GABBRO-DIORITE

CONTACTS OF COMPOSITIONALLY DISSIMILAR ROCKS

The invasion of igneous materials in a sufficient quantity to heat the neighboring country rock to an equivalent temperature establishes chemical gradients in a system which is essentially free from thermal gradients but which has a high temperature. This high temperature encourages reaction in a direction which tends to establish chemical equilibrium in the system.

The contact of Quincy granite with a 2-foot diameter inclusion of Blue Hills porphyry in the Old Colony Crushed Stone Quarry, Quincy, Massachusetts is represented in Figure 3. The zenolith is separated from the enclosing granite by a 3-inch wide transition zone. The granite is composed chiefly of large, angular microperthite grains, quartz, and much strongly pleochroic, dark-green hornblende and riebeckite. The transition zone is mineralogically similar to the granite, but contains a somewhat greater amount of amphibole. A striking feature of the transition zone is the presence of "eyes" of quartz and in places potash feldspar which are surrounded by, or enclose, annular zones of hornblende. Considerable accessory apatite is

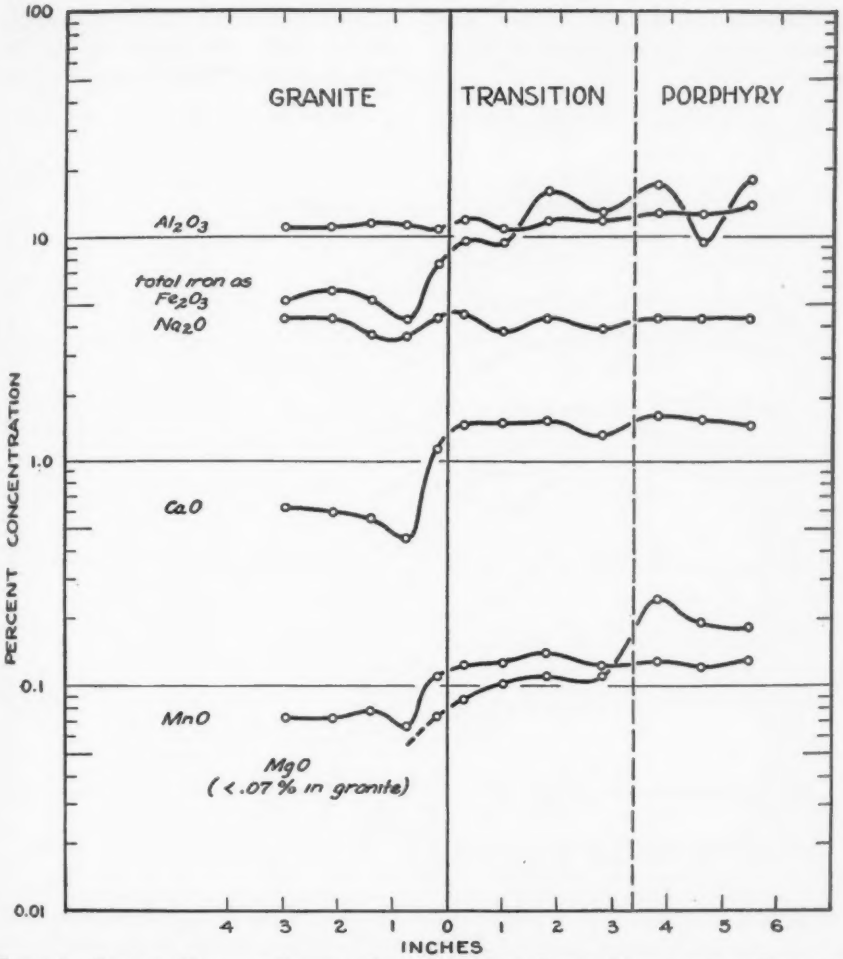

FIGURE 3.—CHEMICAL VARIATION ACROSS A CONTACT BETWEEN QUINCY GRANITE AND AN INCLUSION OF BLUE HILLS PORPHYRY

present. The alteration products are chlorite, sericite, epidote, and calcite. The Blue Hills porphyry was not examined in thin section. It is described by Emerson (1917, p. 192) as composed dominantly of feldspar and augite with accessory quartz, apatite, and magnetite or ilmenite. Thermal gradients in this instance were probably non-existent as the enclosing magma would certainly have heated the xenolith to its own temperature, and the curves, therefore, represent only the effect of chemical gradients. It is very interesting to note that the transition zone is mineralogically similar to the granite, but chemically it is more closely allied to the porphyry.

A second example of the result of compositional differences at high temperatures (Fig. 4) is shown by the chemical variations across a gradational contact between Dedham granodiorite and a rudely spherical, 6-inch diameter, quartzite xenolith. The specimens were obtained from a larger boulder on the southeast shore of Spot Pond, Winchester, Massachusetts.

The additions of alumina, lime, and soda to

the quartzite from the granodiorite is shown mineralogically as feldspathization of the quartzite.

The contact of the Medford diabase dike (300 feet thick) with rhyolite at Pine Hill, Medford, Massachusetts, is represented by

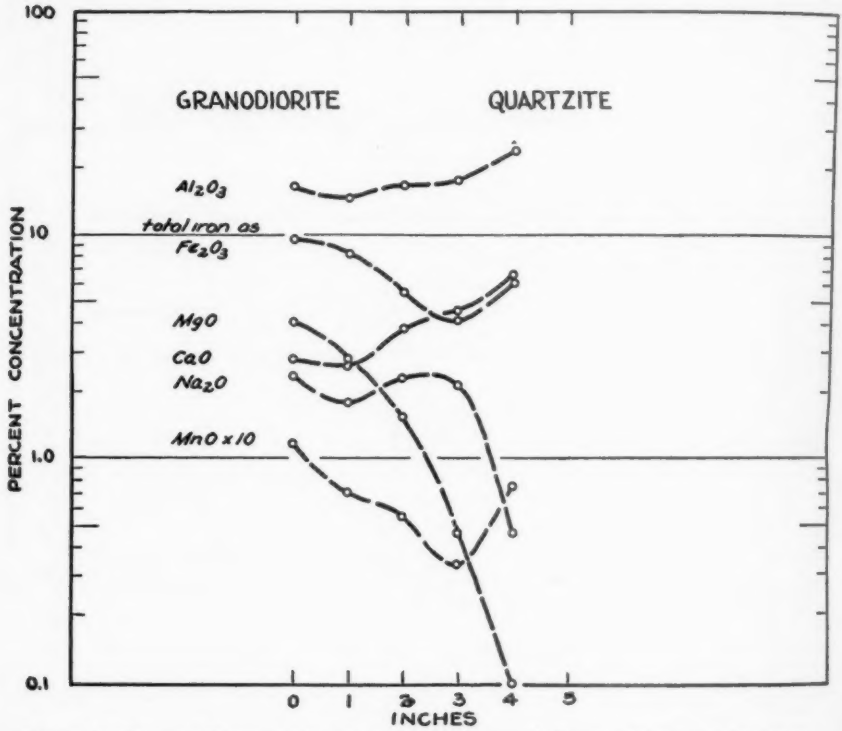

FIGURE 4.—CHEMICAL VARIATION ACROSS A CONTACT BETWEEN DEDHAM GRANODIORITE AND AN INCLUSION OF WESTBORO QUARTZITE

Figures 3 and 4 show that at elevated temperature reaction between compositionally different rocks will produce a rock of intermediate composition. The curves connecting the igneous rocks and their inclusions in Figures 3 and 4 are, however, not smooth, indicating either that reaction was not complete or that hydrothermal solutions of a later magmatic stage have masked the reaction effects.

Dikes, flows, and similar small igneous bodies will usually have contacts across which both thermal and concentration gradients were present. In such a system it is improbable that equilibrium conditions existed, and the irregular nature of the variations across such contacts support this view.

Figure 5. The two points plotted for the analysis of the rhyolite 0.75 of an inch from the contact represent the magnitude of the sampling error. Rhyolite chips were taken at this position in the field and a slab sample was cut from a large hand specimen in the laboratory. The two specimens were prepared and analyzed separately. Migration of iron, magnesia, lime, and manganese from the diabase to the rhyolite is especially well shown by this set of analyses.

The basal contact of the Holyoke diabase flow with Triassic arkosic sediments on the northeast slope of Mt. Holyoke, Massachusetts, is represented in Figure 6.

Two feet above the contact, the flow is a slightly porphyritic, slightly amygdaloidal dia-

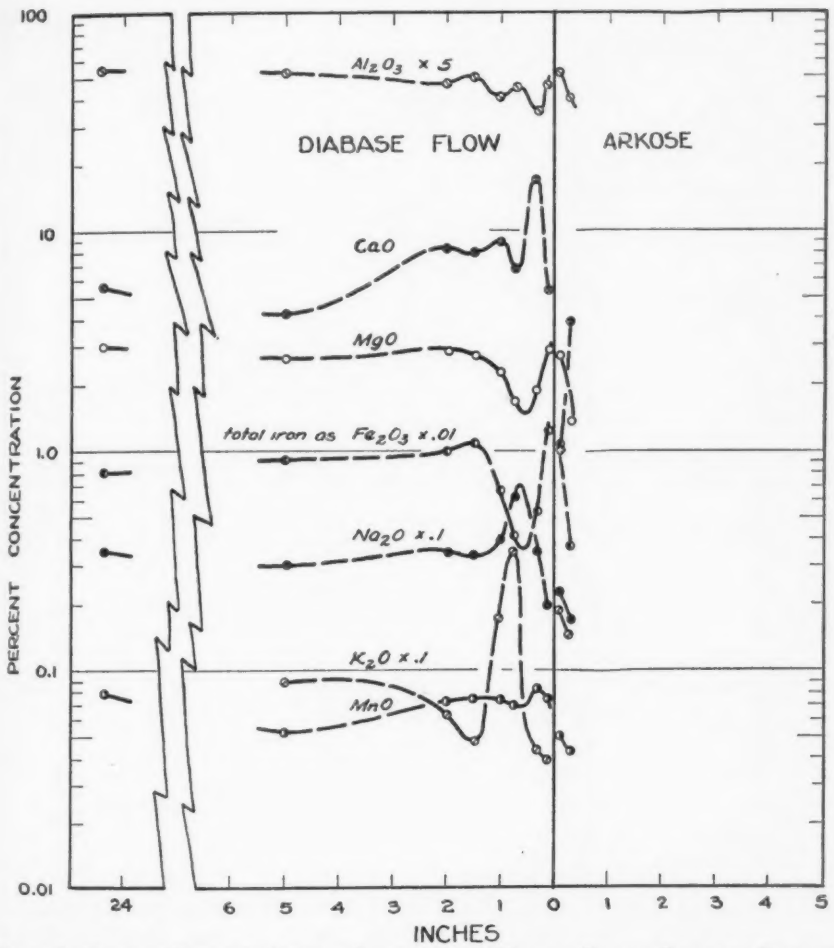

FIGURE 5.—CHEMICAL VARIATION ACROSS A CONTACT OF THE MEDFORD DIABASE DIKE WITH RHYOLITE

base composed of labradorite laths with interstitial augite, magnetite, and green chlorite. The amygdules are infilled with calcite. Six inches above the contact, the grain size is smaller, calcite is found as patchy grains as well as fillings, the chlorite content has increased, and pyroxenes are lacking. Calcite and chlorite continue to increase as the contact is approached, kaolin appears, and the rock assumes an altered appearance. Within an inch of the contact, the flow rock grades into a narrow (a fourth of an inch) bleached zone containing a large amount of kaolin. This zone in turn changes rather sharply into a narrow zone whose chief constituents are calcite and green chlorite.

The underlying arkosic sandstone seems little affected by the overlying flow. It is composed chiefly of detrital quartz grains with some detrital feldspar and muscovite grains. The interstices are filled with red iron oxide and chlorite. The sediments underlying the flow are very thin-bedded, and, consequently, chemical analysis was not carried much beyond the contact.

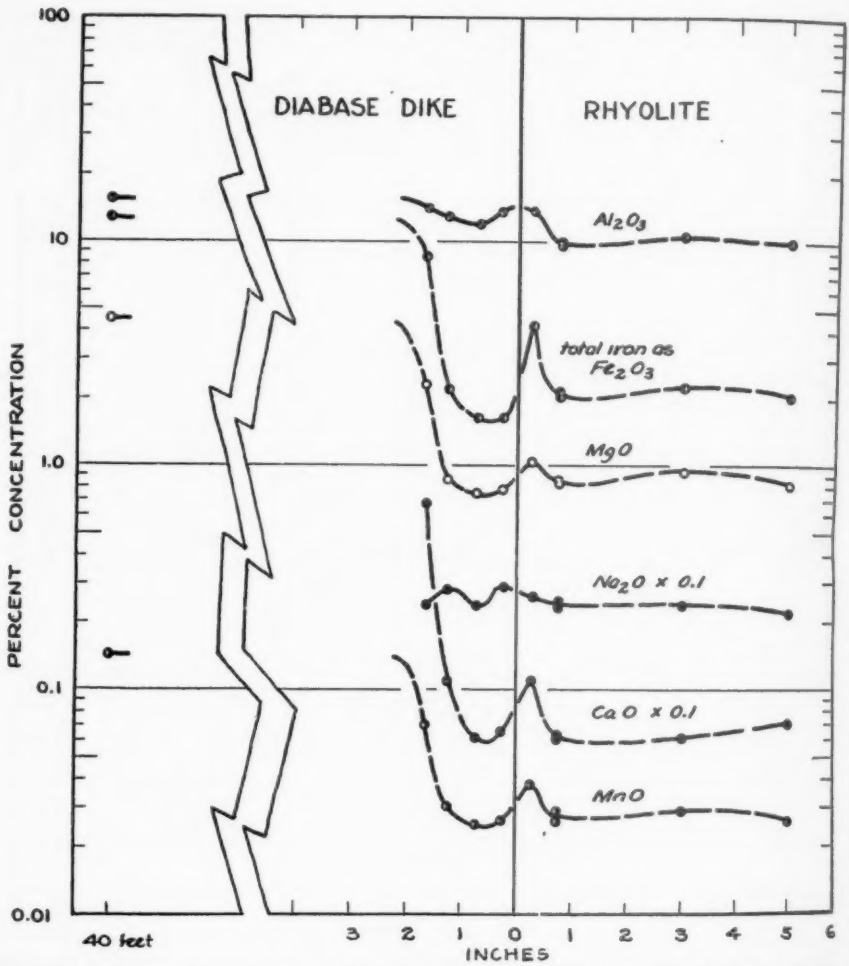

FIGURE 6.—CHEMICAL VARIATION AT THE BASAL CONTACT OF THE HOLYOKE DIABASE FLOW WITH ARKOSE

The two analyses of arkose are both within the bed adjacent to the contact.

CONCLUSIONS

Variation in rock composition near hot contacts is well established by the analyses presented; it is not thought that any major differences from the variation curves will be found by application of more precise techniques.

The majority of the contact zones examined have apparently been modified at a late magmatic stage by aqueous solutions, and it is to these "transfusive" effects that the variation curves would seem to be most directly related. The presence of low-temperature hydrothermal products coupled with the absence of high-grade contact effects in the vicinity of the majority of the contacts investigated is considered as positive evidence that most of the chemical variations observed are due to the movement of material in an aqueous medium.

Two sources of aqueous media are presumably present at contacts; either may be the

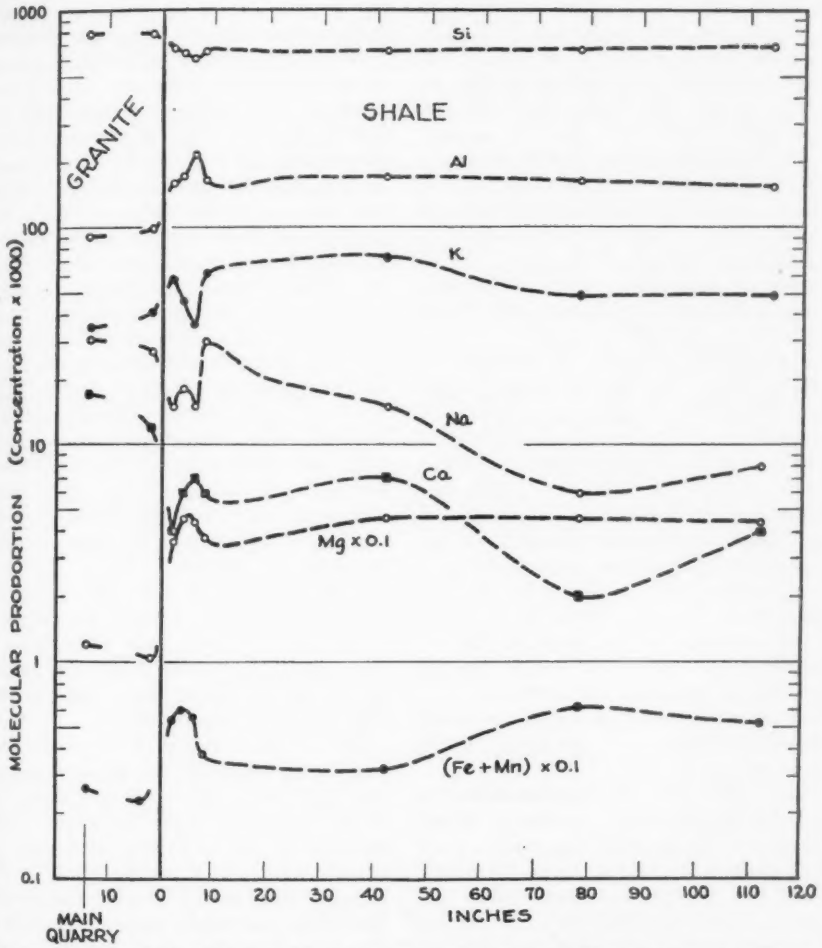

FIGURE 7.—CHEMICAL VARIATION ACROSS THE DARTMOOR GRANITE—BURRATOR SHALE CONTACT

more important in a specific instance. One source is the pore fluid initially present in the country rock. The other is in the partial magma which develops through fractionation of an intrusive igneous body. The character of these aqueous solutions may be highly variable in space and time. The concentration of elements in solution in the pore fluids is chiefly dependent upon the rocks in which they are found. The character of an igneous partial magma depends upon the chemical composition of the initial intrusion, its state of crystallization, temperature, and the containing pressure; its character will vary with time according to the line of crystallization being followed and the degree of loss of the fugitive constituents.

Rapid intrusion and cooling will restrict the possible liquid phases in the magma to the one which is present at the particular stage of fractionation during which emplacement occurred. Small trap dikes may be expected to provide the simplest partial magma situation, or even to have no partial magma sufficiently mobile or long-lived to effect appreciable trans-

fer of material into the walls. The variations associated even with small trap dikes show, however, that transfer has taken place, but whether as a result of emanations from the dike or reaction resulting from heated wall rock pore fluids is not known.

The mobile units present in the various possible fluid phases are most probably hydrated cations. The rate at which these ions may move would seem to be related to the viscosity of the medium, temperature, effective size of the hydrated ion, and valence state. The terminal distance for appreciable ionic migration would be fixed by a threshold energy value (different for different ions) whose spatial location is determined by the thermal gradient. Steep gradients such as those associated with dikes can conceivably cause complex localizations of ions by their movement in time and space.

The direction in which the mobile components will move is believed to depend upon the concentration gradient between two contrasted rock types, or, more properly, the difference in concentration of ions in the pervasive liquid phase. Thermal gradients between intrusive and country rocks may also influence the direction of movement. Parallel chemical and thermal gradients would provide an optimum situation for unidirectional movement. Opposed gradients might set up a complex movement picture.

The distribution of temperature around an intrusion depends upon a number of factors, the chief being the initial temperature of the intrusion and the rate at which this heat is transferred through the adjacent rocks. The transfer of heat from the intrusion is primarily dependent on the rate at which fluids can pass through the rock and secondarily on the conductivity of the rock. Changes in thermal energy might be caused by addition of bulk magma to the intrusive, local exothermic and endothermic reactions, and pressure changes.

The localization of atoms, other than that dictated by thermal considerations, is the result of a complex interplay of solution and precipitation between the pore fluids, mineral solid phases, and continuously changing partial magma diffusing through the solidified portions of the intrusion and the country rocks.

An excellent overall picture of the transfusion process is given by Walker and Poldervaart in their paper on the Karroo Dolerites (1949). They indicate that highly aqueous partial magmas, concentrated by fractional crystallization, escape by diffusion into the wall where they mingle with the pore fluid of the country rock and provide a low viscosity medium through which diffusion can readily take place. The high water content of this medium will promote ionization and assist in the breakdown of some of the complex silicates of the partial magma. The combined sedimentary pore fluid and magmatic emanations seek to establish phasal equilibrium with the solid material through which they circulate and reciprocal reactions follow. In the specific instances of transfusion investigated by Walker and Poldervaart, magnesium and iron seem to have been added to the sediment while silicon was added to the transfusing liquid. Other constituents show irregular variations in accordance with the particular stage at which fractionation took place.

This general idea is upheld by a series of analyses across the contact of the Dartmoor granite with Burrator shale (Brammall and Harwood, 1932, p. 195) which when plotted (Fig. 7) yield curves strikingly similar to those obtained in the present research.

As the contacts analyzed in the present work were chosen because of their igneous character, it is very interesting to see how closely the observed results fit in with the scheme of geochemical changes leading to granitization as outlined by D. L. Reynolds (1946). All the chemical features of Reynolds's scheme have been found reproduced on a minute scale at the contacts studied when appropriate rock types are considered; it would appear that transfusion and granitization are manifestations of a single process.

Quantitative spectrographic techniques have many advantages over some of the accepted analytical methods and may be applied to a large number of geological and geochemical problems.

Outstanding advantages of spectrographic procedure are its accuracy combined with its speed, low cost, and coverage. It is estimated that a complete rock analysis for the major constituents may be obtained in 2 to 3 hours

at a cost of only a few dollars, and some 20 minor constituents may also be determined with a relatively small increase in time and cost.

The accuracy and precision of quantitative spectrographic analysis compares favorably with the results obtained by chemical wet methods, modal analysis, and flame photometry (Dennen, *et al.*, in press). In addition, the errors in spectrographic techniques are well known since multiple determinations are always made, thus allowing the results to be considered from a statistical viewpoint.

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Manuscript Submitted to the Secretary of the Society, November 11, 1950

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 559-564 JUNE 1951

ANALOGY BETWEEN DIFFUSION AND ELECTRICAL CONDUCTIVITY IN POROUS ROCKS

BY L. J. KLINKENBERG

ABSTRACT

Mass transfer in the diagenesis of sediments is accomplished by the movement of solutions, and by diffusion. In a recent paper, Garrels *et al.* (1949) showed that the rate of transport by diffusion is independent of the permeability, but dependent upon an "effective directional porosity." Values of this parameter for several different rocks were determined by diffusion measurements. In view of the analogy which exists between diffusion and electrical conduction this result could have been predicted from electrical data, and the "effective directional porosity" is more easily determinable by electrical measurements.

In an open liquid-filled space the equations for diffusion and for electrical conduction, respectively, are:

$$g = - D_l \operatorname{grad} c,$$
$$i = - K_l \operatorname{grad} V,$$

where D_l and K_l are the diffusion constant and the electrical conductivity of the liquid.

In a rock composed of nonconductive minerals, diffusion and electrical conduction will occur only in the pore space, and along sinuous paths. Hence, in such a case, the effective diffusion constant and electrical conductivity will be given by

$$D_r = \frac{D_l f}{L}$$

$$K_r = \frac{K_l f}{L}$$

where f is the porosity and L a "lithologic factor" depending upon the sinuosity of the path. The geometrical factor f/L is the "effective directional porosity" of Garrels *et al.* It is the same for both electrical conduction and diffusion and can be determined much more easily by electrical than by diffusion measurements.

These deductions are confirmed by experiment and several tables of data for various rocks are presented.

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INTRODUCTION

The diagenesis of sedimentary rocks is partially the result of transport of material through porous rocks. Two mechanisms are generally accepted for this transport:

(a) Transfer of material by movement of solutions through the pores of the rock as a

result of the existence of pressure gradients. The rate of transfer is then determined by the permeability of the formation.

Table 1.—Lithologic Factors of Unconsolidated Sands and Packs of Glass Beads

Formation	Particle size	Porosity	Permeability (10^{-11} cm^2)	Lithologic factor L
white sand	125–210 μ	0.403	22,800	1.60
dune sand	< 297 μ	0.398	24,800	1.64
glass powder	120–840 μ	0.470	—	1.71
glass spheres	8 mm	0.435	14,000,000	1.43
glass beads	72% vol. 8 mm 28% vol. 1 mm	0.262	—	1.57
glass beads + sand	63% vol. 8 mm 25% vol. 1 mm 12% vol. fine sand	0.102	2,500	1.96

(b) Transfer of material by diffusion, as a result of the existence of concentration gradients.

A recent study by Garrels *et al.* (1949) showed experimentally that the rate of transport by diffusion is independent of the permeability but dependent on what the authors call an "effective directional porosity." The purpose of this paper is to show that this result could have been predicted from the analogy which exists between diffusion and flux of electricity. It will further be shown that the "effective directional porosity" can be evaluated from a measurement of the electrical conductivity of a rock saturated with a liquid of known conductivity in a much simpler way than by the measurement of the actual diffusion co-efficient.

Electrical Conductivity of Porous Media

Consider an assemblage of straight capillaries, filled with a liquid the specific conductance of which is K_l; the inner cross sectional area of the capillaries is supposed to be f cm^2 per cm^2. (In other words, the assemblage has a porosity of f). The specific conductance (K_r) of such an assemblage in the direction of the capillaries would be:

$$K_r = K_l \times f. \tag{1}$$

In an actual porous medium, the tortuosity of the pore channels will cause the specific conductance to be smaller:

$$K_r = \frac{K_l \times f}{L} \tag{2}$$

in which L is determined by the lithology of the rock specimen, and may therefore be called the "lithologic factor." Sometimes equation (2) is also written as follows:

$$K_r = \frac{K_l}{F}$$

in which F is called the formation resistivity factor (Archie, 1942).

Equation (2) is only valid, or in other words, L is a constant, if the conductivity of the rock material can be neglected with respect to the conductivity of the liquid with which the rock is saturated.

For unconsolidated sands or packs of glass beads, the lithologic factor varies between 1.4 and 2 (Table 1).

The presence of cementing material always increases the lithologic factor which, dependent upon degree of consolidation, may vary from 2 to 5 and even higher values for consolidated rocks (Table 2).

Analogy between Diffusion Constant and Specific Electrical Conductance[1]

A number of physical processes are governed by analogous laws. For example, the electrical current density is proportional to the potential gradient:

$$i = -K \text{ grad } V \tag{3}$$

in which:

i = current density
K = specific electrical conductance
V = potential.

The diffusion process is governed by a similar law:

$$g = -D \text{ grad } c \tag{4}$$

in which

g = quantity of material passing through unit cross-sectional area per unit time.
D = diffusion constant
c = concentration.

The similarity between the diffusion constant and the specific electrical conductance is apparent from a comparison of equations (3) and (4).

By analogy to equation (2), the apparent diffusion constant (D_r) in a porous medium may be written as:

$$D_r = \frac{D_l \times f}{L} \tag{5}$$

where D_l is the diffusion constant of the material diffusing in the fluid with which the rock is saturated. We might call D_l the "free" diffusion constant.

According to equation (5), it should be possible to calculate diffusion through porous media with the aid of the lithologic factor determined by measurement of the electrical conductivity; or measurements of the diffusion through porous media should lead to the same lithologic factor as that calculated from the electrical resistivity. This will be illustrated with some examples from the literature.

The diffusion of CO_2 through air-saturated sand and clay has been measured by E. Buckingham (1904). (*See* Table 3.) In the last column of Table 3, the lithologic factor has been calculated by means of equation (5) using a value of 0.168 cm²/sec for D_l (the constant for the diffusion of CO_2 in air at 27°C, at which the experiments were carried out).

Table 2.—Lithologic Factors of Consolidated Sandstones

Formation	Porosity	Permeability (10^{-11} cm²)	Lithologic factor
White sandstone.....	0.260	6,050	2.00
Yellow sandstone....	0.241	4,920	2.66
Limy sandstone.....	0.191	190	4.02

The values for the lithologic factor obtained in this way are therefore of the same order of magnitude as the values for unconsolidated material obtained by measurement of the electrical conductivity. The two values marked with an asterisk refer to experiments in which the porosity of the soil-air medium was greatly decreased by the addition of water. Since the rate of diffusion of CO_2 in water is about ten thousand times lower than that of CO_2 in air, high values for the lithologic factor are to be expected. The value found will, moreover, depend on the distribution of the water in the pack.[2]

The second example is derived from measurements of the diffusion of water vapor through unconsolidated sands made by F. Zunker (1930) (Table 4). In the calculation of the lithologic factor, a value of 0.282 for D_l, the diffusion constant of water vapor in air, has been used.

The data of Table 4 show that the lithologic factors found for sands which do not contain liquid are in good agreement with those found by electrical conductivity measurements of similar sands; they are lower than those obtained

[1] The author's attention has been drawn to a publication by Schofield and Dakshinamurti (1948) also dealing with the analogy between diffusion and electrical conductivity.

[2] As the amounts of water added are not specified it is not possible to calculate the actual remaining porosity of the soil-air medium and hence to calculate the "correct" lithologic factors, which will be smaller than the values given in the table.

TABLE 3.—DIFFUSION OF CARBON DIOXIDE THROUGH DIFFERENT KINDS OF SOIL
(From Buckingham, 1904)

Kind and condition of soil	Porosity	D_r (cm²/sec)	L
Dune sand, 4.8 per cent H_2O	0.561	0.0628	1.50
Dune sand, 4.8 per cent H_2O	0.460	0.0358	2.16
Garden loam, air dry	0.485	0.0360	2.26
Garden loam, air dry	0.430	0.0279	2.58
Garden loam, H_2O added	0.312	0.0195	2.69
Garden loam, 18.6 per cent H_2O	0.491	0.0402	2.05
Garden loam, 18.6 per cent H_2O	0.399	0.0299	2.25
Garden loam, 18.6 per cent H_2O	0.319	0.0201	2.67
Cecil clay, 19.6 per cent H_2O	0.475	0.0335	2.38
Cecil clay, 19.6 per cent H_2O	0.351	0.0179	3.30
Cecil clay, 19.6 per cent H_2O	0.249	0.0137	3.04
Cecil clay, 20.1 per cent H_2O	0.463	0.0296	2.63
Cecil clay, 20.1 per cent H_2O	0.347	0.0147	3.97
Cecil clay, H_2O added	0.253	0.0046	9.23*
Cecil clay, air dry	0.609	0.0400	2.56
Cecil clay, air dry	0.520	0.0234	3.72
Cecil clay, air dry	0.468	0.0202	3.89
Windsor sand, subsoil 4.2% H_2O	0.549	0.0570	1.62
Windsor sand, subsoil 4.2% H_2O	0.331	0.0249	2.22
Windsor sand, H_2O added	0.163	0.0388	7.04*

TABLE 4.—DIFFUSION OF WATER VAPOR THROUGH DRY UNCONSOLIDATED SAND PACKS

Type of sand	Porosity	Length of sand column (cm)	Cross-sectional area (cm²)	Vapor pressure gradient (g/cm²/cm. 10^6)	Diffusion constant D_r (cm²/sec)	Lithologic factor L
River sand I	0.3633	25.5	12.78	0.667	0.0729	1.41
River sand II	0.3571	25.5	12.88	0.667	0.0683	1.47
River sand III	0.3602	26.0	12.91	0.654	0.0683	1.48
Hohenbockaer Glass sand	0.3792	26.5	12.56	0.641	0.0628	1.70

TABLE 5.—LITHOLOGIC FACTORS OF LIMESTONES

Limestone	Porosity	Permeability (millidarcys)	Effective directional porosity (%)	Lithologic factor L
Greenhorn	0.340	3–9	10	3.4
Oread	0.114	0.007–0.01	3.1	3.7
Fort Riley	0.182	0.6–0.8	5	3.6

in Table 3 for porous formations containing water.

The third examples refers to the measurements of Garrels *et al.* (1949) who calculated from their diffusion experiments the apparent ratio of core area to core length (A'/l') (calculated from $\frac{A'}{l'} = \frac{V\left(\ln \frac{\Delta c_0}{\Delta c_f}\right)}{2D_t\,\Delta t}$ in which ∇c_0 and Δc_f are, respectively, the differences in concentration between the compartments at the beginning and the end of the time interval Δt and V is the volume of the solution of each compartment) and defined as "effective directional porosity" a quantity: $\frac{A' \times l}{A \times l'}$, in which l and A are the actual length and cross-sectional area of the sample respectively. It follows from our definition of the lithologic factor that:

$$L = \frac{f}{\text{"Effective directional porosity"}}.$$

Table 5 presents the lithologic factors of the samples used by Garrels *et al.*

The data of run 2 (see Garrels *et al.*, 1949, Table 3) have been left out, as the value of the free diffusion constant (D_i) at 25°C has been used in the calculation of the effective porosity for the experiments at 45°C and 65°C.

Finally we would add the results of an experiment where both the diffusion constant and the lithologic factor were measured, the latter by determination of the electrical conductivity. A synthetically consolidated core was connected on one side to a chamber containing a small dish of distilled water and on the other end to a chamber which contained a dish of P_2O_5. The core was mounted in a thermostat maintained at 25°C. It was several days before stationary conditions were obtained, and the rate of water transport could be determined by weighing the P_2O_5-dish before and after a known time interval. The lithologic factor found in this way was 1.9. After that, the same core was saturated with an NaCl solution; the lithologic factor calculated from the electrical conductivity was found to be 1.7. The discrepancy between these values is considered to be within the limits of experimental error.

As, from an experimental point of view, determination of the specific electrical conductance is far simpler than that of the diffusion constant, the correspondence between these two characteristics promises to be of assistance in solving problems where the rate of diffusion of materials through porous media has to be calculated.

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Royal Dutch Shell Laboratory, Amsterdam, Holland

Manuscript Received by the Secretary of the Society, June 8, 1950

121°30′ 25′ 20′

39°55′

50′

0 ½ 1 2 MILES

1951

N 18°

TRUE NORTH MAGNETIC NORTH

Granodiorite

Younger intrusive rocks (late Jurassic?)

Miocene

Older intrusive rocks

Tobin

Storrie

Bucks Creek

Rock Creek

Western Pacific

Bucks

Feather River

North Fork of Feather River

Grizzly Creek

Bear Ranch Hill 4880

Wild Cat Creek

Gravel Range Road

To Quincy

Four Trees

Walker Plain

Marble Creek

Coyote Gap

Pulga

gd mba mb gr ssc ls mr msc hsc qp di g

25′ 20′ 121°15′ 39°55′ 50′

EXPLANATION

Younger intrusive rocks (late Jurassic?)

- b — Basalt — Miocene — TERTIARY
- gr — Granite and trondhjemite — JURASSIC
- gd — Granodiorite
- di — Diorite
- g — Gabbro
- tp — Trondhjemitic porphyry
- mdg — Metasomatic diorite and gabbro

Older intrusive rocks

- pt — Porphyritic trondhjemite
- mg — Metagabbro
- ssc — Serpentine, soapstone and talc schist

- mr, qp — Metarhyolite (mr) and quartz porphyry (qp)
- md, dp — Metadacite (md) and dacite porphyry (dp)
- mba, hp — Metabasalt (mb), amphibolite (a) and hornblende porphyry (hp)
- ag — Agglomerate
- hsc — Hornblende schist
  Bedded tuffaceous sediments metamorphosed in amphibolite facies
- asc — Fine grained amphibole-epidote (chlorite) schist
  Tuffs and tuffaceous sediments metamorphosed in amphibolite or epidote-amphibolite facies

TRIASSIC (?) AND JURASSIC

- ls — Limestone
- q, msc, c — Mica schist with quartzite (q) and conglomerate beds (c)
- csc — Chlorite-epidote schist
  Tuffaceous sediments metamorphosed in green schist facies

CARBONIFEROUS (?) AND YOUNGER

- Andalusite and staurolite
- Garnet
- Migmatitic veins
- Dike, showing dip
- Vertical dike

Granodiorite

Tobin · Storrie · Bucks Creek · Bear Ranch Hill · Four Trees · Coyote Gap · Wild Cat Creek · Grizzly Creek · Rock Creek · Marble Creek · Coldwater Creek · North Fork

BUCKS FORK OF FEATHER RIVER · NORTH FORK OF FEATHER RIVER · WESTERN PACIFIC

WALKER PLAIN · GRAVEL RANGE ROAD · TO QUINCY

50'

45'

Bear Ranch Hill

Big Bar Hill

Four Trees

Merrimac

Granodiorite

Junction House

Mountain House

Brush Creek

Headlands Mill

Big Bend

Poe

Pulga

Coyote Gap

INTAKE DAM

SUROEASE MINE

BIG BEND MINE

FORK OF NORTH

WALKER PLAIN

GRAVEL RANGE ROAD

TO QUINCY

Wild Cat Creek

Marble Creek

Clear Creek

Spoon Ravine

Cloon Creek

Pearine Creek

Little Last Chance Creek

Stony Creek

French Creek

Mosquito Creek

gd mba mb mr msc ssc ls gr mg mdg md tp asc hsc hp qp pt di csc

Metagabbro

Serpentine, soapstone and talc schist

Mica schist with quartzite (q) and conglomerate beds (c)

Chlorite-epidote schist
Tuffaceous sediments metamorphosed in green schist facies

Andalusite and staurolite

Garnet

Migmatitic veins

Dike, showing dip

Vertical dike

Granodiorite

Merrimac

WALKER PLAIN

Bear Ranch Hill

Big Bar Hill

Sky High

Merrimac

Junction House

Mountain House

Deer Park

Milsap Bar

Four Trees

Coyote Gap

Brush Creek

Wild Cat Creek

Clear Creek

Pearine Creek

Coon Creek

Pause Creek

French Creek

Mosquito Creek

Marble Creek

Spoon Ravine

Coldwater Creek

North Fork

South Branch

Little North Fork

Bear Gulch

Last Chance Creek

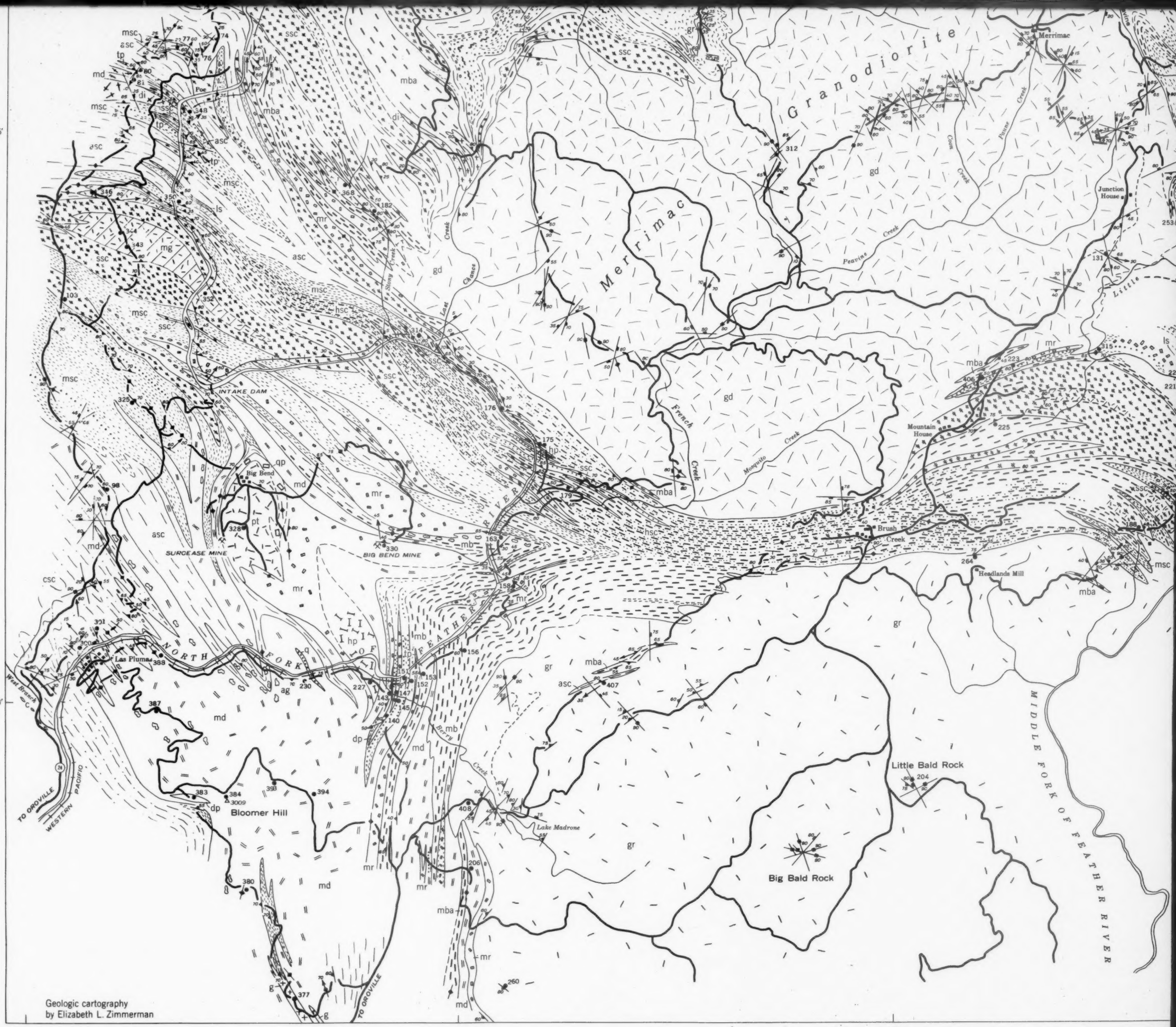
Shoo Fly Creek

GRAVEL RANGE ROAD

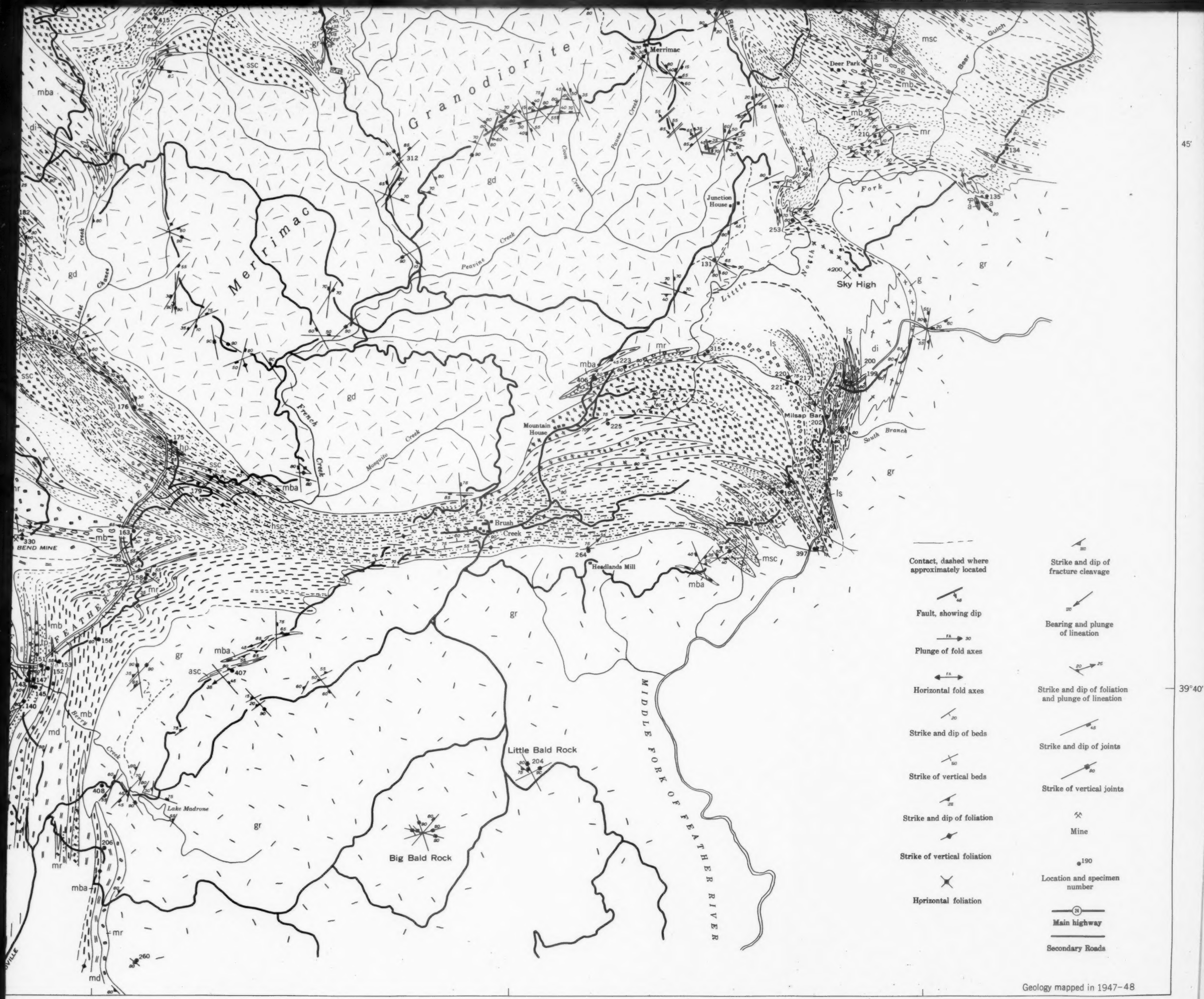
TO QUINCY

FEATHER RIVER

BIG BEND MINE



GEOLOGIC MAP OF THE MERRIMAC AREA, PLUMAS NATIONAL FOREST, CALIFORNIA



Contact, dashed where approximately located

Fault, showing dip

Plunge of fold axes

Horizontal fold axes

Strike and dip of beds

Strike of vertical beds

Strike and dip of foliation

Strike of vertical foliation

Horizontal foliation

Strike and dip of fracture cleavage

Bearing and plunge of lineation

Strike and dip of foliation and plunge of lineation

Strike and dip of joints

Strike of vertical joints

Mine

Location and specimen number

Main highway

Secondary Roads

Geology mapped in 1947–48

45′

39°40′

25′

20′

121°15′

GEOLOGIC MAP OF THE MERRIMAC AREA, PLUMAS NATIONAL FOREST, CALIFORNIA

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 565-608, 21 FIGS., 5 PLS. JUNE 1951

METAMORPHIC AND IGNEOUS ROCKS OF THE MERRIMAC AREA, PLUMAS NATIONAL FOREST, CALIFORNIA

BY ANNA HIETANEN

ABSTRACT

The pre-granitic rocks of an area in the northern Sierra Nevada consist of metamorphosed sedimentary and volcanic series ranging in age from Carboniferous to Jurassic. Synkinematic ultrabasic intrusives, now serpentines, cut these rocks concordantly and discordantly.

Magmatic series ranging from basalt to dacite and soda-rhyolite occur together with the normal basalt-rhyolite series among the meta-volcanics. The younger intrusives (Sierra Nevada series), ranging from gabbros to granodiorites and granites, show great chemical similarity to the meta-volcanic series. Furthermore, soda-rich members are common among the pre-granitic intrusives and younger dike rocks.

The pre-granitic rocks were folded and metamorphosed to green schist and epidote-amphibolite facies prior to emplacement of granodiorite and granite batholiths. The later contact metamorphism affected the areas next to the contacts of the batholitic intrusions, causing crystallization of such minerals as garnet, diopside, epidote, and andalusite.

The plutonic rocks obtained the space needed partly by pushing the country rocks aside and partly by stoping and assimilation.

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Introduction and Acknowledgments

The area investigated lies in the northern Sierra Nevada in the middle western part of the old Bidwell Bar Quadrangle (Fig. 1). The topography is characterized by deep canyons cut by various branches of the Feather River. Highest elevation in the area is 5500 feet above sea level. Canyon walls are steep and rough and may rise as much as 3000 feet from the river. The slopes are thickly vegetated by pines, incense cedar, oak, fir, manzanita, mountain laurel, poison oak, etc. In the higher elevation, pine and fir form the forests and the burned-over areas are frequently covered with a dense growth of manzanita. Many narrow logging roads and abandoned railroads make the area more accessible. Exposures are good in the deep canyons but few and poor at higher elevations.

The work was done on a project grant from The Geological Society of America to which the writer wishes to express her gratitude. Thanks are also due to Mr. Pentti Ojanperä, who performed the chemical analyses in the Institute of Geology at the University of Helsinki, to Dr. Earl Ingerson for critical reading of the manuscript and to Mrs. Stefanie Z. Dobrin, who assisted the author in the field work during one month.

Previous Work

The reconnaissance mapping of the area was done by H. W. Turner (1898).

The main contact between the older rocks and the granite intrusions follows closely the line indicated on this map, but he made no attempt to map the oldest complex nor the granitic rocks in detail. The folio shows the following rock types and gives a short description of each: amphibolites and serpentines (earlier than late Cretaceous in age), Calaveras

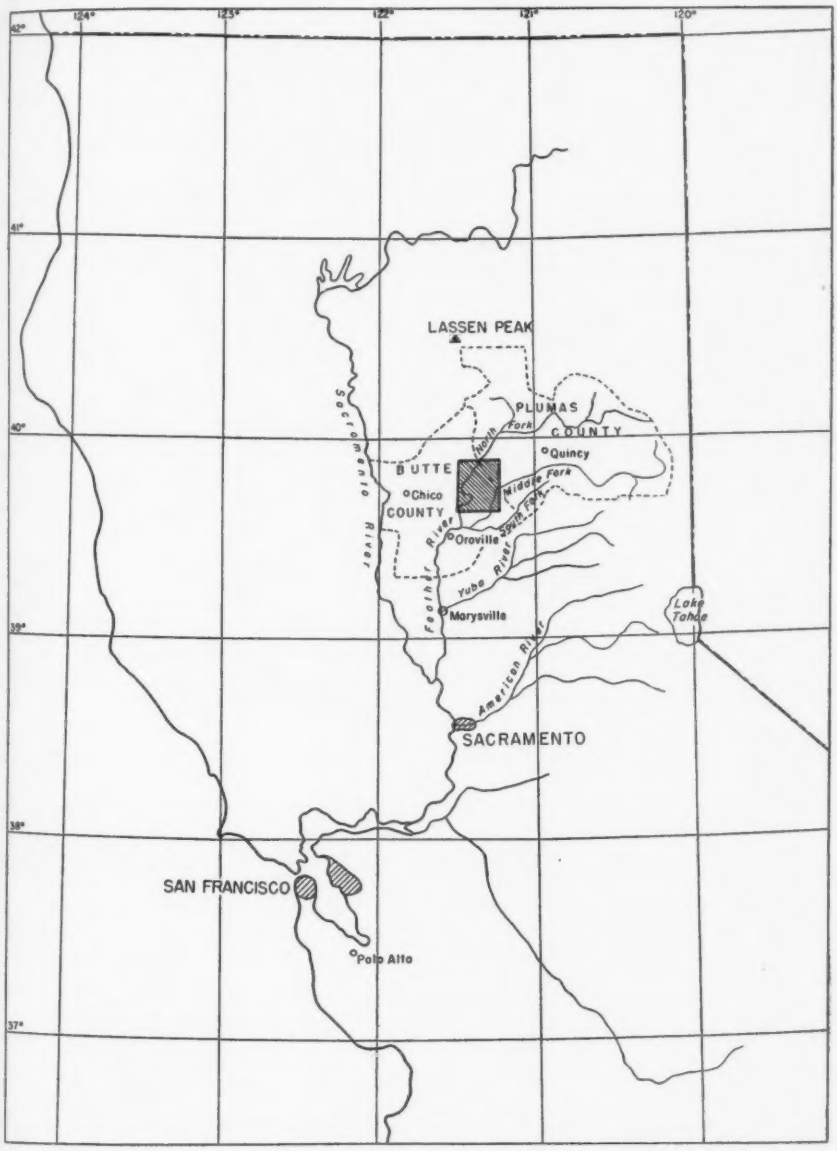

FIGURE 1.—LOCATION OF THE AREA

formation (Carboniferous), intrusive rocks ranging from granite to quartz diorite in composition (earlier than late Cretaceous), and older basalt (Neocene).

PRESENT CLASSIFICATION

All the rocks of the present area, excluding the Walker Plain basalt, belong to the "Bed-

rock Complex" and were involved in the Nevadan Orogeny of the late Jurassic.

The Bedrock Complex is divided into: (1) the older pre-granitic complex and (2) the granitic rocks. The pre-granitic complex consists of various volcanic, plutonic, and sedimentary rocks. The metavolcanics range in composition from basalts and dacites to soda-rhyolites and rhyolites. Some of the metasediments are overlain by metavolcanics, but another part is folded and metamorphosed together with the metavolcanics in such a manner that it is difficult to separate two age groups among these metamorphic rocks. Rather there has been a long period of volcanic activity and contemporaneous sedimentation and then folding and metamorphism. The sediments are mostly mica schists but lenticular bodies of quartzite and limestones are included in them. No unconformities were found between the volcanic and sedimentary rocks. The volcanic rocks include thick layers of agglomerates and tuffs. Sills and dikes of ultrabasic and gabbroic rocks are folded and metamorphosed together with the metavolcanics and metasediments. Ultrabasic rocks were converted to serpentines, soapstones, and talc schists; gabbros to metagabbros of amphibolite facies.

No fossils were found to enable the correlation of the pre-granitic complex with the other series known to occur in the Sierra Nevada.

Intrusives ranging from gabbros and diorites to granodiorites, granites, trondhjemites, and pegmatites are younger than the metamorphosed series. Two types of dikes are distinguished: older dikes occurring only in the older rock series and younger dikes which transect the intrusives as well as the older rocks.

The Walker Plain basalt (Neocene, according to Turner) forms the youngest rock of the area.

Petrographic Description

Metavolcanics

Introduction.—The metamorphosed old volcanics form the main part of the older complex. The sedimentary rocks and tuffs are underlain or intercalated with them. The various layers in many localities are so thin that it has been impossible to present them in the scale of the map (Pl. 1), therefore the layering as shown on the map is symbolic in these localities. The comparison of Figure 5 with the map shows this clearly. The structure is highly lenticular. It is best shown on the map as lenticular occurrence of the agglomerates and rhyolites.

The following four rock types are distinguished among the metavolcanics: (1) metabasalts (2) metadacites (3) metarhyolites, in part rich in soda (4) metamorphosed tuffs and agglomerates.

Metabasalt.—Metabasalts are most widely spread among the metavolcanics. They are dark, schistose or massive, fine- to medium-grained, rarely coarse rocks. Abundant amygdules which are drawn out into oval lenticles are common. Blocky weathering is typical. In some localities, such as the northern part of the Marble Creek gorge, the old basalt is only a little deformed and looks like the younger undeformed basalt at Walker Plain, just west of this area. The resemblance is emphasized by the empty oval vesicles from which the calcite filling has weathered out. The only difference to be seen with the naked eye is the occurrence of fine recrystallized hornblende needles oriented at random in the older metamorphosed basalt.

The mineralogical composition of metabasalts is rather uniform. Light-green hornblende and epidote are the main constituents. Quartz and plagioclase are the light minerals and their amount varies considerably. Magnetite is a common accessory. Biotite, calcite, and chlorite may occur in minor amounts. Relicts of former phenocrysts are common (Fig. 2). The vesicles are usually filled with quartz and epidote. Plagioclase phenocrysts contain abundant small hornblende needles.

The refractive indices and the color of the hornblende vary. The common type in the fine-grained epidote-amphibole rocks shows (sp. no. 5): $\gamma' = 1{,}658 \pm 0.001$, $\bar{a}' = 1{,}646 \pm 0.001$, $\gamma \wedge C = 21°$, and pleocroism: $\gamma = \beta$ = bluish green, α = pale green.

These properties indicate a hornblende in which Fe:Mg is about 20:80. The fine-grained, schistose metabasalt no. 156, here chosen to represent the most basic member of the extrusive series, contains green hornblende needles with $\alpha = 1.685 \pm 0.001$ and $\gamma \wedge C = 15°$.

TABLE 1—CHEMICAL AND MINERALOGICAL COMPOSITION OF METAMORPHIC AND IGNEOUS ROCKS, THEIR NIGGLI VALUES AND NORMS

	Metamorphosed extrusive rocks						Older intrusives		Rocks of metasomatic origin								Int
Anal. No.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	
Loc.	Uralite porphyrite Brush RR Station, Mouth of Berry Creek	Metabasalt RR Cut E off Mouth of Berry Creek	Metamorphosed dacite tuff 2 miles E of Las Plumas	Metadacite Bloomer Hill	Metarhyolite Brush Creek	Metamorphosed soda rhyolite Berry Creek	Altered trondhjemite Big Bend	Serpentine 1 miles S of Pulga Bridge	Serpentinized peridotite Milsap Bar Road	Metasomatic gabbro S of Pulga	Metasomatic gabbro S of Pulga	White vein material S of Pulga	Calculated 50% No. 1 + 50% No. 12	Gabbro 0.5 mile south of Rock Creek	Inclusion in diorite 118 Grizzly Creek	Diorite Grizzly Creek	Granodiorite Grizzly
Loc. No.	175	156	230	384	406	147	328	2	188	28A	28C	28B		270	117	118	1
SiO_2	47.76	49.12	47.40	53.58	68.22	73.92	62.62	42.79	46.52	45.87	45.70	41.11	44.43	47.12	51.88	58.38	
TiO_2	0.63	1.35	0.74	0.80	0.56	0.34	0.39	0.00	0.13	0.12	0.21	0.06	0.34	1.05	1.12	0.81	
Al_2O_3	14.85	16.72	18.19	17.04	14.83	12.76	18.00	1.87	1.81	24.03	26.89	31.79	23.32	11.01	17.91	16.22	
Fe_2O_3	1.92	3.84	2.88	2.15	0.08	1.36	2.80	3.75	5.39	2.47	0.80	0.56	1.24	2.80	2.56	1.45	
FeO	5.89	7.69	7.47	6.90	3.88	2.01	1.29	3.66	3.52	1.07	2.37	0.93	3.41	7.61	6.32	4.88	
MnO	0.11	0.17	0.13	0.20	0.11	0.07	0.11	0.10	0.11	0.06	0.06	0.00	0.05	0.11	0.16	0.13	
MgO	13.58	5.58	5.48	4.19	1.61	0.35	1.04	36.68	27.52	8.44	5.90	2.67	8.12	13.76	4.80	4.27	
CaO	9.09	10.39	12.25	9.43	0.86	2.48	4.89	0.10	8.26	15.19	14.13	19.09	14.09	10.87	9.17	7.21	
Na_2O	3.84	3.43	2.34	2.58	3.97	4.71	4.88	0.00	1.33	1.50	1.64	1.08	2.41	3.84	3.73	3.59	
K_2O	0.46	0.58	0.57	0.46	4.30	0.93	1.89	0.00	0.08	0.41	0.98	0.28	0.37	0.84	1.20	1.21	
P_2O_5	0.34	0.17	0.15	0.11	0.07	0.31	0.04	0.03	0.00	0.00	0.00	0.00	0.17	0.14	0.37	0.66	
CO_2	0.00	0.00	0.00	0.00	0.00	0.00	0.54	0.00	0.71	0.00	0.00	0.00	0.00	0.00	0.00	0.63	
H_2O^+	1.79	0.62	2.75	2.33	0.99	0.41	1.78	10.97	4.82	1.03	1.65	2.62	2.20	1.18	1.04	0.49	
H_2O^-	0.06	0.05	0.08	0.08	0.23	0.05	0.12	0.22	0.14	0.14	0.14	0.20	0.13	0.04	0.05	0.06	
	100.32	99.71	100.43	99.85	99.71	99.70	100.39	100.17	100.34	100.33	100.47	100.39	100.28	100.37	100.31	99.99	
quartz			7.8	14.85	27.47	38.87	17.49								3.3	16.7	
orthoclase	2.37	2.85			12.70	1.00									4.7	2.3	
plagioclase	An_{28} 30.40	An_{29} 29.50	An_{3} 16.6	An_{28} 23.40	An_{3} 33.85	An_{11} 44.50	An_{8} 40.20			An_{28} 49.5	An_{28} 54.5	An_{28} 14.2		An_{28} 10.0	An_{28} 50.5	An_{27} 45.9	An_{2}
muscovite			6.2	5.20	8.00	1.00	19.00					1.0		9.7		0.4	
biotite					17.34	8.10									5.0	9.9	
chlorite	10.23		10.0				6.70			1.0	2.4	2.0					
hornblende	33.74	46.14	19.5	22.70					62.3	39.2	32.7	10.1		76.8	31.6	20.3	
epidote		16.62	36.3	29.60		4.40	15.33			10.3	10.3	Zoisite 72.7				0.4	
calcite							1.23		1.6							0.7	
apatite	0.57	0.40	0.4	0.26	0.17	0.68	0.09							0.3	0.8	1.0	
titanite			1.6											1.6		0.5	
rutile							0.31										
ilmenite	1.00	4.21		1.22					7.1								
magnetite	2.00		2.0	2.00		1.25								2.0	4.1	1.9	
hematite									serp + ol 29.0						H_2O 0.3		
rest				water 0.61													
	100.31	99.72	100.4	99.84	99.53	99.80	100.35		100.0	100.0	99.9	100.0		100.4	100.3	100.0	
si	96.7	115.3	109.1	163.8	308.0	395.0	235.0	70.96	78.5					91.5	126.0	176.0	2
ti	0.97	2.4	1.28	1.64	1.9	1.38	1.1		0.16					1.53	2.05	1.83	
qs	−36.00	−19.5	−15.3	33.4	88.24	183.8	45.8	−29.04	−30.5					−41.7	−16.8	24.5	
al	17.74	23.1	24.7	27.3	39.55	40.3	39.83	1.8	1.8					12.6	25.6	28.75	
fm	54.27	42.0	38.7	27.5	26.34	17.6	18.13	98.0	81.05					56.5	35.4	35.07	
c	19.81	26.2	30.5	27.5	4.17	14.3	19.72	0.2	14.90					22.6	28.3	23.31	
alk	8.18	8.7	6.1	7.6	29.94	27.8	22.30		2.25					8.3	10.7	12.87	
k	0.07	0.10	0.14	0.106	0.42	0.115	0.20		0.04					0.13	0.17	0.18	
mg	0.76	0.47	0.49	0.45	0.41	0.16	0.32	0.99	0.85					0.70	0.49	0.55	
c/fm	0.36	0.62	0.79	0.73	0.16	0.81	1.09	0.002	0.18					0.40	0.80	0.66	
										0.18						12.30	
qu				8.75	21.45	36.50	16.40			2.40	5.80			4.96	7.10	7.20	
or	2.74	3.46	3.35	2.74	25.50	5.53	11.10		0.5	13.80	10.15			22.10	31.70	30.45	
ab	27.12	29.22	19.82	21.92	33.70	40.00	41.30		8.9	57.50	63.00			10.13	28.60	24.40	
an	21.91	28.46	37.40	33.48	4.27	10.85	21.00	0.3			1.94			5.07			
ne	3.05								15.3	7.49	2.98			18.90	6.21	1.42	
wo	8.90	9.25	9.40	5.27	2.05	0.35	0.35	1.8									
cor									2.1						10.26	10.59	
ac							2.58	46.3	32.2	1.34					6.87	7.04	
en		7.82	8.53	10.40	3.99	0.87		1.7	1.1								
hy		4.96	6.45	9.95	6.32	2.14		31.4	25.4	13.80	10.15			24.10	1.16		
fo	23.85	4.36	3.75					1.6	1.1		2.67			7.50	0.80		
fa	6.06	3.01	3.02				0.09	0.1						0.31	0.81	1.46	
ap	0.74	0.37	0.35	0.26	0.16	0.68	0.74		0.3	0.23	0.40			2.00	2.13	1.53	
il	1.20	2.56	1.40	1.52	1.03	0.65	3.35	5.5	6.8	3.29	1.16			4.05	3.70	2.11	
mt	2.78	5.57	4.16	3.13	0.11	1.97	0.47			0.21						1.43	
hem							1.23		1.6								
cal							1.78	11.3	4.96		1.79			1.18	1.04		
H_2O	1.85	0.67	2.83	2.41	1.22	0.41											
	100.30	99.71	100.46	99.83	99.80	99.95	100.39	100.0	100.26	100.24	100.04			100.30	100.38	99.93	9
Classification according to Johannsen	3212 andesite	3212 andesite	238 quartz basalt	238 quartz basalt	217 sodaclase rhyodacite	228 dacite	228 tonalite		428 harzburgite					3212 meladiorite		228 tenalite	227 granodi
Classification according to Shand	Ab > An Soda-basalt	Ab > An Soda-basalt	An > Ab > Or lime-dacite (Mesotype, metaluminous)	An > Ab > Or lime-dacite Mesotype, metaluminous)	Ab > Or > An Soda-rhyolite (leucocratic peraluminous)	Ab > An > Or soda-dacite (leucocratic peraluminous)	Ab > An > Or soda-tonalite		peridotite					Ab > An > Or soda-gabbro		Ab > An > Or soda-tonalite (mesotype, metaluminous)	Ab > An soda-tona (leucocr metalu nous)

AL COMPOSITION OF METAMORPHIC AND IGNEOUS ROCKS, THEIR NIGGLI VALUES AND NORMS

Rocks of metasomatic origin | Intrusives of Sierra Nevada Series

9	10	11	12	13	14	15	16	17	18	19	20	
Serpentinized peridotite Milsap Bar Road	Metasomatic gabbro S of Pulga	Metasomatic gabbro S of Pulga	White vein material S of Pulga	Calculated 50% No. 1 + 50% No. 12	Gabbro 0.5 mile south of Rock Creek	Inclusion in diorite 118 Grizzly Creek	Diorite Grizzly Creek	Granodiorite Grizzly Creek	Granodiorite Grizzly Creek	Granite-Trondhjemite Bald rock	Pegmatite Grizzly Creek	
188	28A	28C	28B		270	117	118	119	121	204	116	
46.52	45.87	45.70	41.11	44.43	47.12	51.88	58.38	66.78	68.49	72.97	75.92	SiO_2
0.13	0.12	0.21	0.06	0.34	1.05	1.12	0.81	0.45	0.43	0.13	0.08	TiO_2
1.81	24.03	26.89	31.79	23.32	11.01	17.91	16.22	14.56	14.57	15.04	13.05	Al_2O_3
5.39	2.47	0.80	0.56	1.24	2.80	2.56	1.45	0.80	1.12	0.40	0.48	Fe_2O_3
3.52	1.07	2.37	0.93	3.41	7.61	6.32	4.88	2.87	2.01	0.79	0.43	FeO
0.11	0.06	0.06	0.00	0.05	0.11	0.16	0.13	0.09	0.06	0.04	0.03	MnO
27.52	8.44	5.90	2.67	8.12	13.76	4.80	4.27	2.27	1.92	0.11	0.17	MgO
8.26	15.19	14.13	19.09	14.09	10.87	9.17	7.21	4.67	3.97	2.65	1.10	CaO
1.33	1.50	1.64	1.08	2.41	3.84	3.73	3.59	3.59	3.38	5.12	3.14	Na_2O
0.08	0.41	0.98	0.28	0.37	0.84	1.20	1.21	2.24	2.54	2.04	5.23	K_2O
0.00	0.00	0.00	0.00	0.17	0.14	0.37	0.66	0.60	0.56	0.46	0.00	P_2O_5
0.71	0.00	0.00	0.00	0.00	0.00	0.00	0.63	0.50	0.27	0.00	0.00	CO_2
4.82	1.03	1.65	2.62	2.20	1.18	1.04	0.49	0.34	0.45	0.37	0.33	H_2O^+
0.14	0.14	0.14	0.20	0.13	0.04	0.05	0.06	0.07	0.07	0.08	0.07	H_2O^-
100.34	100.33	100.47	100.39	100.28	100.37	100.31	99.99	99.83	99.84	100.20	100.03	
						3.3	16.7	27.7	30.8	30.7	35.8	quartz
						4.7	2.3	8.4	9.6	9.9	29.0	orthoclase
	An_{25} 49.5	An_{25} 54.5	An_{10} 14.2		An_{33} 10.0	An_{33} 50.5	An_{17} 45.9	An_{11} 42.8	An_{23} 39.8	An_{17} 52.0	An_{13} 30.2	plagioclase
			1.0		9.7		0.4		1.0	2.0	2.0	muscovite
						5.0	9.9	9.9	9.9	2.5	2.0	biotite
	1.0	2.4	2.0					0.1	0.1			chlorite
62.3	39.2	32.7	10.1		76.8	31.6	20.3	6.8	4.0			hornblende
	10.3	10.3	Zoisite 72.7				0.4	1.2	2.6	1.2		epidote
1.6							0.7	0.9	0.6		0.5	calcite
					0.3	0.8	1.0	0.9	0.4	1.1		apotite
					1.6		0.5	0.5	0.3			titanite
												rutile
7.1												ilmenite
serp + ol 29.0					2.0	4.1	1.9	0.7	0.7	0.5	0.5	magnetite
										0.2		hematite
						H_2O 0.3						rest
100.0	100.0	99.9	100.0		100.4	100.3	100.0	99.9	99.8	100.1	100.0	
78.5					91.5	126.0	176.0	266.8	295.5	376.0	467.0	si
0.16					1.53	2.05	1.83	1.34	1.4	0.5	0.37	ti
−30.5					−41.7	−16.8	24.5	88.0	110.9	148.0	209.60	qz
1.8					12.6	25.6	28.75	34.30	37.0	45.75	47.25	al
81.05					56.5	35.4	35.07	23.85	23.45	7.56	6.15	fm
14.90					22.6	28.3	23.31	20.15	18.40	14.69	7.25	c
2.25					8.3	10.7	12.87	19.70	21.15	32.00	39.35	alk
0.04					0.13	0.17	0.18	0.29	0.33	0.21	0.52	k
0.85					0.70	0.49	0.55	0.52	0.52	0.11	0.25	mg
0.18					0.40	0.80	0.66	0.78	0.78	0.20	1.18	c/fm
	0.18						12.30	24.20	28.70	30.02	34.77	qu
0.5	2.40	5.80			4.96	7.10	7.20	13.30	15.30	12.65	31.00	or
8.9	13.80	10.15			22.10	31.70	30.45	30.50	28.60	43.35	26.60	ab
	37.50	63.00			10.13	28.60	24.40	16.35	14.75	10.22	5.46	an
		1.94			5.07							ne
13.3	7.49	2.98			18.90	6.21	1.42					wo
								0.15	0.86	0.54	0.21	cor
2.1												ac
32.2	1.34					10.26	10.59	5.63	4.76	0.27	0.42	en
1.1						6.87	7.04	4.75	2.18	0.99	0.32	hy
25.4	13.80	10.15			24.10	1.16						fo
1.1		2.67			7.50	0.80						fa
					0.31	0.81	1.46	1.30	1.21	0.99		ap
0.3	0.23	0.40			2.00	2.13	1.53	0.85	0.82	0.24	0.15	il
6.8	3.29	1.16			4.05	3.70	2.11	1.16	1.62	0.58	0.70	mt
	0.21											hem
1.6							1.43	1.14	0.61			cal
4.96		1.79			1.18	1.04		0.34	0.45	0.45	0.40	H_2O
100.26	100.24	100.04			100.30	100.38	99.93	99.87	99.86	100.30	100.03	
428 harzburgite					3212 meladiorite		228 tonalite	227 granodiorite	227 granodiorite	127 leucogranodiorite	127 leucogranodiorite	
peridotite					Ab > An > Or soda-gabbro		Ab > An > Or soda-tonalite (mesotype, metaluminous)	Ab > An > Or soda-tonalite (leucocratic, metaluminous)	Ab > An > Or soda-tonalite (leucocratic, metaluminous)	Ab > Or > An soda-granite	Or > Ab > An potash-granite	

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Winchell's tables give the relation Fe:Mg = 70:30 for this mineral. The chemical and mineralogical composition of this metabasalt is shown under No. 2 in Table 1. A fine-grained, massive, basic rock near the mouth of Berry Creek (sp. 230) is chemically very similar but differs considerably mineralogically. It contains more epidote and less amphibole than rock No. 156. The plagioclase in rock No. 230 consists of albite (An_5); but in No. 156 of oligoclase (An_{16}). The amphibole in No. 230 is a pale tremolitic variety with $\gamma' = 1.655 \pm 0.001$, $\gamma \wedge C = 14°$. Pale chlorite is another dark constituent. Rock No. 230 probably represents a metamorphic tuff layer. Locations of these two specimens as well as the others mentioned are indicated by large numbers on the map (Pl. 1).

Metadacites. — Metadacites form the main part of the high Bloomer Hill south of Big Bend. The belt of these intermediate metavolcanics extends southward, trending about S. 5°–10° E., and northwest to the Big Bend area. The last-mentioned extension is well exposed at the road cuts north from the West Branch bridge. The rock is fine- to medium-grained, gray, and very hard. Part of it shows clearly a fragmental texture. Blocky weathering is common. Between the hard layers there are schistose fine-grained layers, many of them showing fragmental texture and some containing amygdules of quartz and epidote.

The minerals in the metadacites are plagioclase (An_{10}), quartz, hornblende, and epidote in varying amounts. The rock at Bloomer Hill and in its surroundings contains abundant plagioclase phenocrysts and quartz - epidote amygdules which make the general appearance on the weathering surface fairly light greenish-gray (Pl. 3, fig. 1). The plagioclase phenocrysts range usually from 2 to 4 mm. in length. In a road cut 1½ miles south of the Bloomer Hill lookout, however, euhedral phenocrysts ranging from 5 to 15 mm. in diameter abound.

The plagioclase phenocrysts contain numerous inclusions of epidote and sericite. In many instances, these minerals make up 70 per cent of the phenocrysts and oligoclase only about 30 per cent. Aggregates of chlorite and epidote, or of tremolite, chlorite, and epidote show straight boundary lines and probably represent altered phenocrysts of pyroxene or amphibole. The chemical analysis of a dacite (No. 384) with few amygdules is shown in Table 1 (No. 4). About 4 per cent of the quartz is contained in the amygdules. After subtracting that from

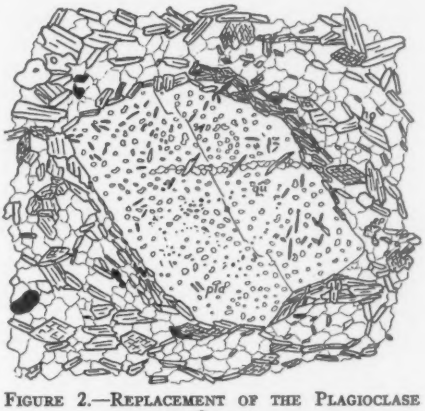

FIGURE 2.—REPLACEMENT OF THE PLAGIOCLASE BY QUARTZ

Quartz veinlets and blebs (qu) replace plagioclase (An_{34}) (pla) in uralite porphyry No. 249. Gravel Range road. 30×.

the total quartz, the rock still contains enough free quartz to be called dacite.

The fine-grained gray variety in the Big Bend area and at Las Plumas shows pyroclastic textures. Agglomerates with round bombs ranging from 5 to 15 cm. and more in diameter are well exposed along the road from Bloomer Hill to Las Plumas. (Pl. 2, fig. 1). The bombs consist of amygdaloidal fine-grained or porphyritic meta-dacite. The minerals in the fine-grained groundmass are epidote, hornblende, quartz, and plagioclase.

Some dacite flows near Las Plumas contain considerable chlorite in addition to tremolite and epidote and thus show a lower degree of metamorphism. Plagioclase laths are then replaced by albite, calcite, and epidote. Amygdules consist of calcite, which readily weathers away at the surface, giving the rock the appearance of a recent lava flow. A few augite phenocrysts, completely altered to chlorite, hornblende, and epidote, and some quartz phenocrysts were found in the same rock. (Loc. 394).

Metamorphosed soda-rhyolites. — Light-gray silicic porphyritic rocks, here called soda-rhyo-

lites, are found in a belt east of the mouth of Berry Creek, in the Big Bend area, and in a belt north of Walker Plain. Some of the rocks discussed under this heading are true extrusives but others are sill or dike rocks cooled near the surface. Because of recrystallization it is impossible to tell whether the groundmass was originally glassy. Pyroclastic textures indicating an extrusive nature are preserved, as at the mouth of Berry Creek (loc. 152) and north of Marble Creek camping ground (loc. 282). On the other hand, rhyolite occurs in cross-cutting dikes just west of locality 147. Two minor lenticles occur in the Stony Creek area as shown on the map (Pl. 1). The rock at the mouth of Berry Creek is hard and dark brownish-gray. The size and amount of phenocrysts vary. Some well preserved outcrops contain blue phenocrysts of quartz ranging from 2 to 5 mm in diameter and light-brownish plagioclase (An_{18}) phenocrysts of the same size. The ground mass is fine-grained and granoblastic, consisting of quartz, plagioclase, greenish-brown biotite, with $\gamma = \beta = 1.646 \pm 0.001$, epidote and some chlorite and sericite. The quartz phenocrysts are euhedral but often broken and show undulatory extinction (Fig. 3). Plagioclase phenocrysts contain numerous small inclusions of epidote and sericite and some biotite. Magnetite and apatite are the accessories.

FIGURE 3.—METAMORPHOSED SODA-RHYOLITE
Phenocrysts are quartz (qu) and plagioclase (pla). Dark mineral is biotite. Chlorite (chl) occurs occasionally as an alteration product after biotite. Mouth of Berry Creek, Loc. 147. 21×

The chemical analysis of this rock is shown in Table 1 (No. 6). The amounts of SiO_2, MgO, and FeO are about the same as in the normal rhyolites, but the amounts of K_2O and Na_2O differ strikingly from those in normal rhyolites, in the rock No. 147 the amount of Na_2O is about 5 times that of K_2O. The amount of CaO is larger than that in normal rhyolites. In these respects, the chemical composition is near that of dacites but there is more quartz and less femic minerals in this rock. According to Johannsen's classification (1931), it would belong to the tonalite family and should be called dacite. According to Shand's classification (1947) it would belong to the dacite group and would be called soda-dacite. Comparison of the chemical analysis of the present rock with that of the trondhjemites indicates closest similarity.

The mineral assemblages of both are the same except for a certain amount of epidote present in No. 147, due to the metamorphism in the epidote-amphibolite facies. Rock No. 147 is thus an extrusive equivalent of the trondhjemites. The metadacites of the Bloomer Hill area (Anal. 4, table 1) and the metabasalts north east of it (Anal. 2 table 1) belong to the same series: basalt → dacite → soda-rhyolite.

The author feels that soda-rhyolite is a more suitable name for extrusives which correspond to trondhjemites than dacite or soda-dacite. Among the European petrographers the name quartz-keratophyre has been used for the metamorphic equivalents of soda-rich paleovolcanic rhyolites. However, Rosenbusch (1898) realized that the soda-rhyolite north of Berkeley, California, described by Palache (1893) is identical with the quartz-keratophyres and he called it a neovolcanic quartz-keratophyre. In the western part of the United States, the name keratophyre has been used for soda-rich equivalents of andesites and trachytes, and quartz-keratophyres for soda-rich dacites and rhyolites (A. Knopf, 1918; 1921; 1924; 1929) even if they were fairly unaltered. The occurrence of albite phenocrysts is often considered reason enough for naming the rock a keratophyre. The name soda-rhyolite will there-

fore indicate the petrologic character and genesis of the present rock better than quartz-keratophyre.

In the belt north of Walker Plain, metarhyolites are found interlayered with metabasalts. These metarhyolites are white to bluish-gray, mostly fine-grained, hard, and very little deformed. As a rule they contain more potash feldspar than the corresponding effusives in the southern part of the area; for instance in meta-rhyolite No. 304, potash feldspar is the main constituent of the rock. Phenocrysts are quartz and orthoclase, and the fine-grained groundmass orthoclase, albite, and quartz. Biotite is a scanty additional mineral.

In a road cut along the way from Wildcat to Bear Ranch Hill (Loc. 288 on map), abundant pink euhedral quartz phenocrysts, ranging from 2 to 15 mm in diameter, occur in schistose white metarhyolite. Microscopic examination shows that this rock is hydrothermally altered to a sericite schist. The quartz phenocrysts show some granulation, and sericite occurs along the fractures. The feldspar phenocrysts are completely altered to sericite and have lost their euhedral shape. The ground mass consists of small granoblastic quartz and sericite. Small tourmalines occur along planes of schistosity. Small prisms and grains of rutile occur abundantly in certain small spots, which probably represent remnants of some ferromagnesian mineral, apparently biotite. It is impossible to tell how much of the potash is primary and how much was introduced hydrothermally into this rock.

In the area southeast of Merrimac the amount of potash feldspar in metarhyolites varies. Light-colored layers southwest of Milsap Bar (Loc. 190) and northwest of Deer Park (Loc. 210) consist of metamorphic soda-rhylites while the light-gray metarhyolite northwest from Milsap Bar (Loc. 221) contains both feldspars, and the amount of plagioclase is about twice that of the potash feldspar. This flow is well exposed in the gorge of the Little North Fork. As the other rhyolites also, this flow is very little deformed. Phenocrysts of quartz and feldspar are well preserved. Flow structure can be observed in the fine-grained ground mass which consists of quartz, plagioclase, orthoclase, biotite, and muscovite. Magnetite occurs as an accessory. Estimated percentages of the minerals in this rock are: quartz 39 percent, plagioclase (An_6) 37 percent, orthoclase 18 percent, muscovite 3 percent, biotite 3 percent.

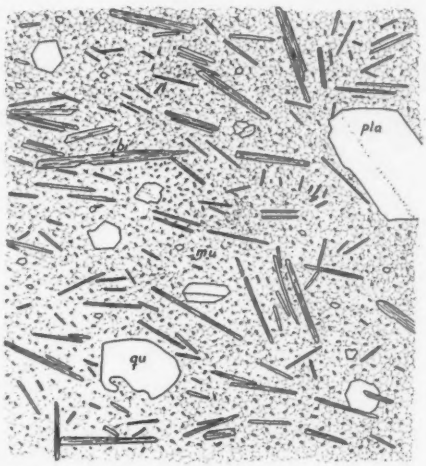

FIGURE 4.—METARHYOLITE RICH IN BIOTITE
Abundant pseudo-hexagonal biotite (bi) plates in metarhyolite No. 406. Phenocrysts are plagioclase (pla) and quartz (qu). 2½ miles north of Brush Creek. 32×

A peculiar rhyolite was found north of Brush Creek Ranger Station. Two 3 meter thick layers of this rock are exposed in a road cut along the old Quincy-Oroville road 2½ miles north of Brush Creek (Loc. 406) and along a logging road half a mile east of this locality (Loc. 223) near the contact of Merrimac granodiorite. This rhyolite differs strikingly from the other rhyolites in the area. It is exceptionally rich in biotite, which occurs in hexagonal plates ranging from 1 to 3 mm. in diameter (Pl. 2, fig. 2). Quartz and plagioclase (An_8) phenocrysts are few and small. Figure 2 of Plate 3 shows that quartz phenocrysts are later than the biotites. The groundmass is fine-grained and granoblastic like that of the other rhyolites. The orientation of the biotite plates shows a good eddy structure like that produced by turbulent flow. Figure 4 shows the structure perpendicular to the platy structure and to the axis of the eddies. E. B. Knopf (Knopf and Ingerson, 1938, page 132) has explained that

this type of flow structure will result if two flows with different velocities come together. In the present area, the relation between the two flows exposed is not clear since there is a younger granite intrusion between them.

The amount of potash (Table 1, No. 5) is greater than that of sodium as in normal rhyolites but, because of the high content of biotite, the amount of orthoclase is only about a third of that of plagioclase and it all is in the fine-grained groundmass.

Schistose layers of volcanic origin.—The fine-grained schistose layers near Las Plumas and south of Bloomer Hill (Loc. 380) are most likely metamorphosed tuffs. They contain abundant chlorite in addition to light-green hornblende; calcite, epidote, albite, and quartz are the other constituents. This mineral assemblage indicates a lower degree of metamorphism than in the corresponding layers in other parts of the area.

Many schistose layers show structures of deformed agglomerates or tuffs, Loc. 200 (Pl. 4, fig. 3), 213 (Pl. 2, fig. 3), 225 (E. of Mountain House), 315. The drawn-out bombs in agglomerate 213 consist of schistose dacite with groups of square plagioclase phenocrysts and fine-grained groundmass, in which the light-green hornblende needles are well oriented parallel to the schistosity. The fine-grained groundmass between the bombs contains more quartz and less hornblende than the bombs. The former outlines of augite can be seen in several large individual hornblende crystals, which have grown over the original boundary lines. Epidote, biotite, chlorite, and magnetite are the additional constituents. In the dacitic tuff (No. 315, Little North Fork), plagioclase occurs in two generations: the old phenocrysts are euhedral and altered to sericite and epidote. The newly crystallized large grains are fresh and include numerous small rounded quartz grains. As a rule the relict phenocrysts are common in tuffaceous layers and provide a good indication of the origin. The distinct layering also is most likely due to a tuffaceous origin (Pl. 3, fig. 3).

The fine-grained, homogeneous schists, consisting mainly of hornblende and epidote with some quartz and plagioclase, are the most common rock types in the area and probably were originally basic tuffs (e.g. Nos. 76, 153, 350).

The harder layers which consist of the above-described metamorphosed lavas are interbedded

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Plate 2.—Biotite-rich metarhyolite and polished agglomerate specimens

Figure 1. Agglomerate with Rounded Bombs

Note quartz amygdules in the bombs. Loc. 387, N. side of Bloomer Hill.

Figure 2. Biotite Rich Metarhyolite Parallel to Lamellar Structure

Loc. 406, 2½ miles N. of Brush Creek. 2×

Figure 3. Drawn-Out Bombs in a Dacite Agglomerate East of Deer Park

Shows the shortest and longest dimension (right) and cross section (left).

Plate 3.—Photomicrographs

Figure 1. Spherical Quartz Amygdules in the Dacite Form About 50% of the Rock

Epidote occurs in the center of the large amygdule. Sp. 388, Las Plumas. + Nicol. 8×

Figure 2. Cross Section of Pseudo-Hexagonal Biotites Showing a Flow Structure

Metarhyolite No. 223, 2½ miles N. of Brush Creek. + Nicol. 8×

Figure 3. Relict Phenocrysts of Plagioclase and Hornblende and a Distinct Layering in a Dark Amphibolite South of Pulga

Younger quartz vein cuts across the layering. Loc. 31.—Nicol. 12×

Figure 4. Contact Facies of the Bald Rock Pluton

Large euhedral to subhedral plagioclase includes small epidote and sericite, and is corroded by quartz. These large crystals as well as the small dull grains with inclusions originate in older rock. + Nicol. 12×

Figure 5. Trondhjemite Porphyry No. 182

Plagioclase appears in lathshaped phenocrysts and hornblende needles in clusters. Stony Creek. + Nicol. 12×

Figure 6. Biotite-Albitite Dike No. 307

Marble Creek. Albite laths show trachitoid texture. Biotite is the only dark mineral. + Nicol. 27×

FIGURE 2

FIGURE 3

FIGURE 1

BIOTITE-RICH METARHYOLITE AND POLISHED AGGLOMERATE SPECIMENS

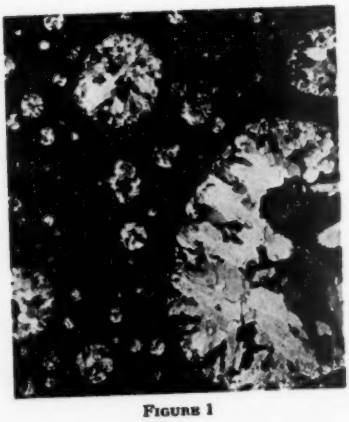

FIGURE 1

FIGURE 2

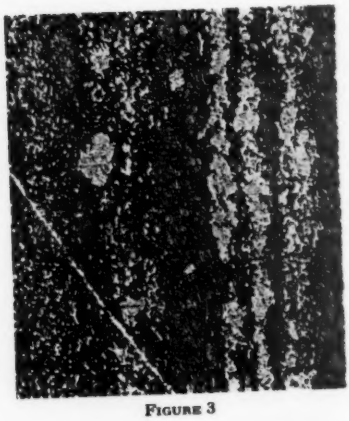

FIGURE 3

FIGURE 4

FIGURE 5

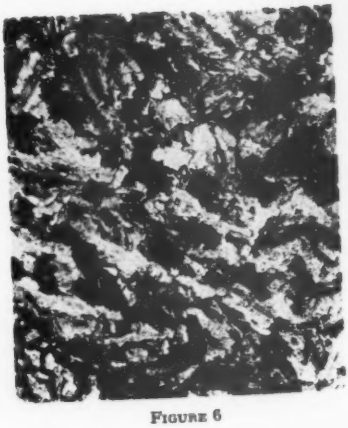

FIGURE 6

PHOTOMICROGRAPHS

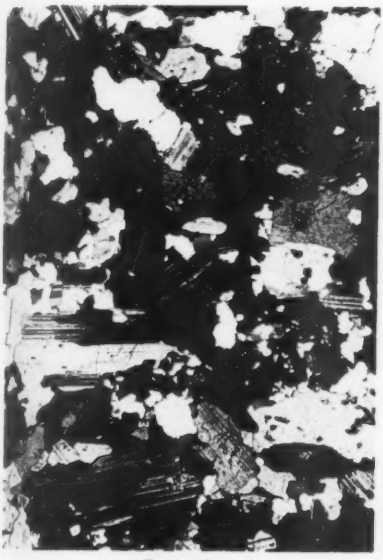

FIGURE 1

FIGURE 2

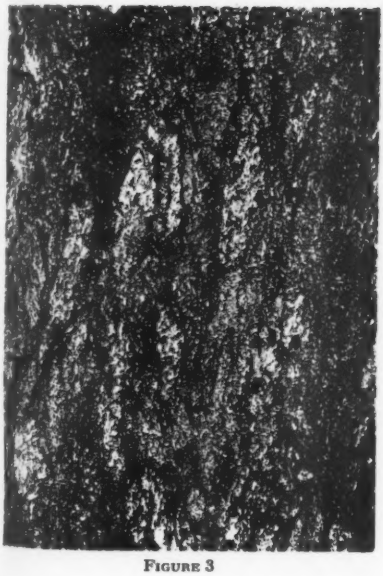

FIGURE 3

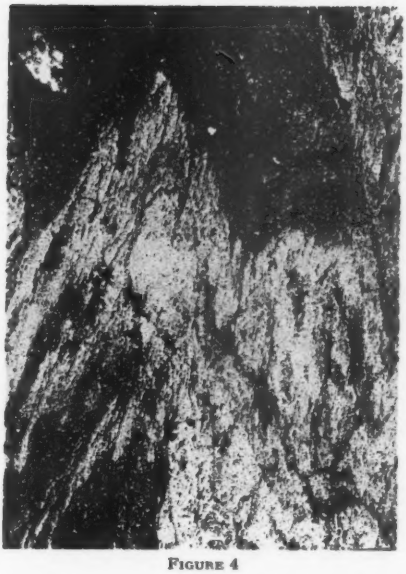

FIGURE 4

PHOTOMICROGRAPHS

FIGURE 2

FIGURE 1

FIGURE 4

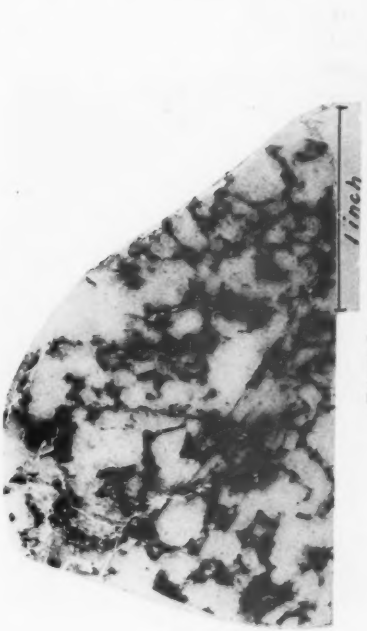

FIGURE 3

CONTACT BRECCIA AND METASOMATIC ROCKS

with these fine-grained schistose rocks. In some localities, such as east of the mouth of Berry Creek (No. 152), these layers contain small lenticles consisting of calcite, epidote, and quartz. The lenticles are probably deformed amygdules and the layer represents a fine-grained vesicular part of an andesitic or dacitic flow. The next layer (No. 151) consists of fine-grained, hornblende-epidote schist in which there are some long, thin lenticles made up of small quartz grains. Large, light lenticles in the same rock consist of small grains of quartz, plagioclase, hornblende, and epidote and therefore have the same composition as the dacites. These lenticles may well represent cinders, and the layer could be a metamorphosed lapilli tuff. A light-gray to white layer, which is interbedded with rhyolite at locality 143, consists of fine-grained, quartz-plagioclase-muscovite-epidote schist. The amount of epidote, about 30 per cent, would raise the percentage of CaO to about 7 per cent. This is considerably higher than the corresponding percentage in the rhyolite. The layer is most likely a dacite tuff in origin.

Another light-gray banded dacite tuff layer occurs as a lenticular inclusion in granite about 2 miles north of Lake Madrone. Plagioclase phenocrysts are partly euhedral, somewhat rounded and include hornblende needles with random orientation. The amount of hornblende and quartz varies in different layers. Hornblende is oriented parallel to the bands.

Schistose layers of soda-rhyolitic composition are common in areas where well preserved, almost undeformed soda-rhyolites occur. Some of these layers appear to be fine-grained schistose soda-rhyolites (No. 145) and some like No. 158 represent tuffs. These metatuffs are fine-grained, gray muscovite-biotite schists with a few euhedral quartz and plagioclase phenocrysts. Partial chloritization of biotite is common as is alteration of plagioclase to epidote and sericite.

The light-gray schistose layer in the metarhyolites in the Marble Creek area (No. 320) differs in its mineral assemblage from the tuffaceous layers described above. The main constituent in this layer is albite (An_7) which makes up about 70 per cent of the rock. Quartz occurs in subordinate amounts (about 20 per cent). Biotite and muscovite and some orthoclase are the additional constituents. Part of the albite occurs in long laths. Small grains of magnetite are well scattered throughout the rock.

Another light-gray, somewhat schistose layer just north of the Marble Creek camping ground (Loc. 282) was probably originally a welded rhyolite tuff. It consists of about 38 per cent

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PLATE 4.—PHOTOMICROGRAPHS

FIGURE 1. DIORITE NO. 118

Plagioclase is anhedral to subhedral and shows no zoning. Hornblende is anhedral and includes quartz and plagioclase. Grizzly Creek. + Nicol. 13×

FIGURE 2. GRANITE NO. 204

Large quartz and microcline grains include zoned, euhedral plagioclase. Little Bald Rock. + Nicol 13×

FIGURE 3. PYROCLASTIC STRUCTURES IN A SCHISTOSE HORNBLENDE-BIOTITE SCHIST

Loc. 200, Milsap Bar.—Nicol. 13×

FIGURE 4. BEDDING, SMALL FOLD AND TRANSECTING CLEAVAGE IN THE BIOTITE SCHIST AT MARBLE CREEK

Loc. 362.—Nicol. 13×

PLATE 5.—CONTACT BRECCIA AND METASOMATIC ROCKS

FIGURE 1. GRANITIC VEINLETS CUTTING THE AMPHIBOLITE AT ANY ANGLE

Showing the starting brecciation near the contact of granodiorite at the mouth of Grizzly Creek.

FIGURE 2. CONTACT BRECCIA JUST BELOW THE AMPHIBOLITE SHOWN IN FIGURE 1

Note the large recrystallized hornblende crystals (black).

FIGURE 3. METASOMATIC DIORITE TWO MILES SOUTH OF PULGA

Plagioclase (white) is subhedral. Hornblende occurs in long prisms. Loc. 28.

FIGURE 4. METASOMATIC GABBRO AS TRANSITIONAL ROCK BETWEEN THE DIORITE IN FIGURE 3 AND THE FINE-GRAINED AMPHIBOLITE

Loc. 28, 2 miles south of Pulga.

quartz, 33 per cent plagioclase (An_8), some orthoclase, sparse biotite, hornblende, and muscovite. Small cubic and irregular magnetite grains abound. Zircon with pleocroic haloes occurs as an accessory. Under the microscope, small lenticular areas show a simultaneous extinction of tiny albite laths with fringed boundaries. The length of these lenticles varies from 1 mm to 2 cm and even more.

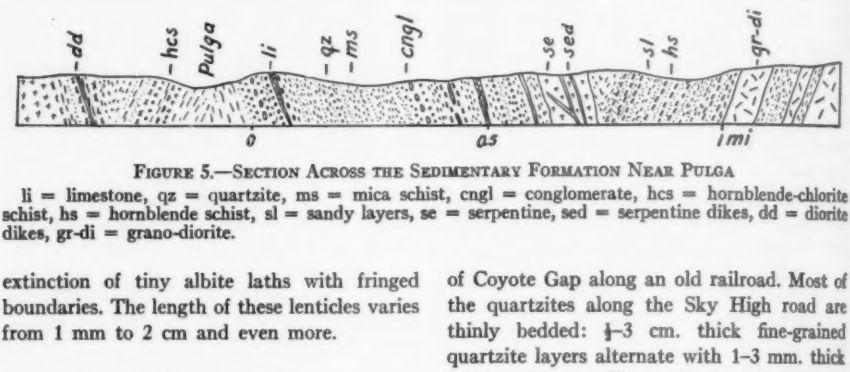

FIGURE 5.—SECTION ACROSS THE SEDIMENTARY FORMATION NEAR PULGA

li = limestone, qz = quartzite, ms = mica schist, cngl = conglomerate, hcs = hornblende-chlorite schist, hs = hornblende schist, sl = sandy layers, se = serpentine, sed = serpentine dikes, dd = diorite dikes, gr-di = grano-diorite.

Metasediments

Introduction. — Metasediments are intercalated and folded with the metavolcanics. Their mode of occurrence seems to indicate that in part they are older than the metavolcanics and in part contemporaneous. Petrographically they belong to the following main groups: quartzites, mica-schists, chlorite-epidote schists, limestones, and various amphibole schists.

Quartzites and quartzite-conglomerates.—The quartzites are fine- to medium-grained hard rocks which show a distinct bedding. The amounts of micaceous minerals, commonly biotite and muscovite, and their distribution in different layers are variable. Chlorite, apatite, and magnetite are the common accessories. Most of these quartzites are light gray or brownish but some, like No. 217, contain thin dark layers which are rich in dust-like magnetite. The pronounced bedding in quartzites west of Milsap Bar is caused mainly by the layers of pure quartzite alternating with layers containing magnetite and biotite. The dark layers are only about 0.3 mm. thick, while the thickness of the light layers varies from 0.3 mm. to 1 cm. (No. 217). A pebbly, biotite-bearing quartzite is exposed in the river under the Milsap Bar bridge (Loc. 202). The light quartzite bed, about 10 m. thick and located 1½ miles southeast of Big Bar Lookout, is composed of one-half to 1 cm. thick layers of fine- to medium-grained quartzite alternating with 1 mm. thick micaceous layers. Another thicker quartzite bed is exposed 1½ miles east of Coyote Gap along an old railroad. Most of the quartzites along the Sky High road are thinly bedded: ½–3 cm. thick fine-grained quartzite layers alternate with 1–3 mm. thick micaceous layers. Fine-grained quartzite lenses in the dacites east of Las Plumas and north of the West Branch bridge were probably cherts originally. The material in the quartzites at the mouth of Berry Creek was derived mainly of rhyolitic and dacitic tuffs.

There are some conglomeratic layers in the tremolite quartzites north of Pulga Railroad Station. These quartzite conglomerates and quartzites are interbedded with mica schists which contain calcite lenticles. Microscopic study shows that the pebbles in the conglomerates consist of quartz and tremolite. Very fine-grained magnetite occurs as an accessory. The similarity of the composition in the pebbles and matrix indicates that these conglomerates are more likely tectonic than sedimentary.

Mica schists.—The distribution of the mica schists is shown on the map (Pl. 1). The occurrences near Pulga (Fig. 5) and those in the Big Bend, Marble Creek, and Milsap Bar areas are the most extensive. The schists near Pulga and in the Big Bend area (Loc. 325) are mostly fine-grained biotite-quartz schists with some muscovite. Those in the Marble Creek and Milsap Bar areas are coarser and darker.

Light greenish-brown to brown biotite is the most common micaceous mineral in all of these mica schists. Muscovite is abundant in some layers, and chlorite is found especially

in the southwestern part of the area. Dust-like magnetite is locally abundant in thin dark layers which alternate with light layers free of this mineral. Quartz veinlets and lenticles are common in the schists. White lenticles ranging from one-half to 4 mm. thick and from 1 to 10 mm. long occur in the dark-gray biotite schist (Loc. 202) exposed under the bridge at Milsap Bar. These lenticles are small, drawn-out quartzite pebbles, and the bed was originally a laminated pebbly sand layer. A few larger, rounded feldspar grains are found in the same rock. Calcite and chlorite occur as minor constituents.

Mica schists along the Sky High road (Loc. 134) are fine-grained, rich in quartz, and show shiny greenish blue-gray cleavage planes. Microscopic examination shows that about 70 per cent of the rock is quartz and 30 per cent is muscovite with some magnetite which is mostly altered to hematite.

Garnet-bearing biotite schists occur 100 m. east of the junction to Bear Ranch Hill Lookout (Loc. 127), 1¼ miles southeast of Big Bar Lookout, a mile east of Coyote Gap (Loc. 126), and south of Walker Plain near the contact of the granodiorite (Loc. 128). Anhedral garnets ranging from 2 to 5 mm. in diameter are embedded in fine-grained biotite schist. Many of them are altered to muscovite and biotite along the cracks and borders. Some layers contain in addition abundant epidote.

The fifth extensive mica schist belt occurs 1¼ miles southeast of Big Bar Lookout (Loc. 415). These schists contain abundant andalusite. In the fine-grained beds, small andalusite grains form 3–7 mm. long oval aggregates. In the coarse schist, the individual andalusite grains range from 0.5 to 2 cm long. Thin section study shows that in certain layers andalusite was crystallized in huge holoblasts which include quartz, biotite, and magnetite as helicitic inclusions while other layers contain groups of small grains. The layers between those rich in andalusite consist of biotite, quartz, some muscovite, staurolite, and magnetite. Staurolite occurs as small grains between biotite and muscovite scales. Its boundary lines, as well as those of muscovite are indistinct. Where andalusite grains appear, there is no more staurolite nor muscovite, but recrystallized clear biotite scales and abundant small magnetite grains are included in andalusite. There is no doubt that andalusite is later than staurolite and that it crystallized at the expense of

FIGURE 6.—ANDALUSITE SCHIST

Staurolite (st) surrounded by andalusite (and) in the andalusite-biotite schist 1¼ miles southeast of Big Bar Lookout, Loc. 415. The other minerals are qu = quartz, bi = biotite, mt = magnetite. 29×.

staurolite and muscovite. The two following reactions have taken place:

(1) $K_2Al_4(OH)_4(Si_6Al_2O_{20}) + 6x[2Al_2SiO_5 \cdot Fe(OH)_2]$
Muscovite 6 Staurolite

$+ \ SiO_2 \longrightarrow K_2Fe_6(OH)_4(Si_6Al_2O_{20})$
2 quartz Biotite

$+ \ 14Al_2SiO_5 + 6H_2O$
14 Andalusite 6 Water

(2) $2Al_2SiO_5 \cdot Fe(OH)_2 \longrightarrow 2Al_2SiO_5 + FeO + H_2O$
Staurolite 2 Andalusite

Only one larger unaltered staurolite grain was found in the thin sections studied, and it was surrounded by andalusite (Fig. 6). Andalusite is commonly found in rocks which have undergone a high temperature metamorphism without shearing stress, as in contact metamorphic rocks. Staurolite is considered to be a stress mineral (Harker, 1932). In this area, the intrusion of granite caused the rise of temperature after the sediments were folded and metamorphosed in a somewhat lower temperature field. The staurolite was crystallized during the folding and the andalusite during the intrusion period.

Chlorite-epidote schists with conglomeratic beds. —The fine-grained schists near Las Plumas and West Branch bridge consist of chlorite, muscovite, epidote, and quartz with some small tremolite needles, sparse biotite, and fine-grained magnetite. The quartz grains are angular or rounded. Certain layers are colored brownish by abundant hematite. Abundance of chlorite and epidote in many layers is due to the high content of magnesium and calcium and indicates presence of basic tuffaceous material in these layers.

TABLE 2.—ANALYSIS OF THE LIMESTONE No. 199, MILSAP BAR

Insoluble	4.68 %
$Al_2O_3 + Fe_2O_3$	0.39
FeO	0.29
MnO	0.014
CaO	52.12
MgO	0.83
CO_2	41.54
	99.864

Conglomeratic layers are intercalated with these schists just south of West Branch bridge, near Las Plumas (Loc. 300) and south of the Intake Dam. The size of the pebbles varies in different layers. Near the West Branch bridge, they range from 2 to 5 cm. long, while in the conglomerate layers near Las Plumas and south of the Intake Dam they are only 7 to 10 mm. long. The small pebbles consist of chert, sericite, quartzite, vein quartz, andesite, and altered plagioclase porphyry. The groundmass contains angular and rounded quartz grains and fragments of ferromagnesian minerals altered to sericite and chlorite, all embedded in fine-grained sericite quartzite.

Limestones.—Limestone occurrences are few and rather small. Largest lenticles are found in the Marble Creek district and near Milsap Bar. Two beds east of Milsap Bar are interbedded with metabasalts and agglomerates. The third one is situated between metabasalts and mica schists and extends farther south than the others. There are no fossils which would indicate the age of these limestones. They are coarse marbles, in which light-gray and white bands alternate. The minerals are calcite, with very little muscovite, pyrite, and quartz. The chemical analysis of a limestone, No. 199, east of Milsap Bar is given on Table 2. It shows that 93.1 per cent of the rock is calcium carbonate. The indices of refraction of this calcite are: $\epsilon = 1.491 \pm 0.001$, $\omega = 1.652 \pm 0.001$.

The mode of occurrence and association may give some clue to the origin of various lenticles. Those which occur as lenticles in the volcanic series may well be hot spring deposits in origin, whereas the limestones interbedded with the meta-sediments are of sedimentary origin.

The amphibolites in a belt about 1 mile north of Deer Park contain numerous calcite lenticles ranging from 1 to 20 mm. broad and 1 to 10 cm. long. In some places about 50 per cent of the rock consists of calcite.

In the same zone, there are several larger lenticles of white to gray marble (Pl. 1). About 350 meters north of Pulga, layers of hornblende-biotite schist, containing white calcite lenticles, are interbedded with mica schist. Calcite is fine- to medium-grained and includes diopside crystals ranging from 1 to 2 cm. in diameter. The diopside is light greenish-gray with a refractive index, $\beta = 1.690 \pm 0.005$.

Tremolite-biotite schists. — Tremolite-biotite schists are interbedded with mica schists and represent dolomitic sand layers, e.g. Loc. 22, 52, and 68. Tremolite occurs as colorless long prisms, many of which are oriented at random and gathered into clusters. Epidote, calcite, and magnetite are the common accessories. These layers are generally rather thin, ranging mostly from 3 to 10 cm. thick and therefore impossible to indicate on the map.

Some layers consist of almost pure tremolite (No. 55) or of tremolite and epidote with some magnetite and quartz. Rutile is a common accessory. A light-gray compact 5 cm. thick cherty looking layer in a mica schist exposed in a road cut 2 miles south of Pulga (Loc. 69) is a very fine-grained tremolite-epidote rock which contains small tremolite-epidote quartzite lenticles in certain layers. The length of the lenticles varies from 0.5 to 15 mm. and the thickness from 0.1 to 2 mm. This bed was most likely a dolomitic clay layer interbedded with pebbly layers of the same composition.

Chlorite may occur instead of or together with biotite in the fine-grained gray hornblende-bearing schists of lower metamorphic facies (Fig. 7). Light-green hornblende instead of tremolite occurs together with biotite as a dark mineral in many thin dark layers, which are intercalated with the mica schists along the Little North Fork (Loc. 220).

Hornblende schists.—Banded rocks consisting mainly of quartz, plagioclase, and hornblende in varying amounts are here called hornblende schists. In the field, they are distinguished from the amphibolites by their lighter color and a distinct banding. Mineralogically they differ from the amphibolites in the high content of quartz. The banding is represented by alternation of dark- and light-gray layers ranging from 1 mm. to 4 cm. thick. The light layers contain up to 70 per cent quartz. The dark mineral, hornblende (about 20 per cent), is a common green variety with the following pleochroism: γ = blue-green, β = green, α = yellowish-green. Plagioclase is present in small amounts in dark layers, but is lacking in many light layers. Biotite occurs in some layers together with hornblende. Epidote and magnetite are the accessories. The banding is most likely due to the bedding of a formation, in which sedimentary and tuffaceous layers alternate. The bedding planes are straight, and beds of even thickness over fairly great distances show that the deposition probably took place in a basin containing water. The major belts of these hornblende schists occur north of Pulga Railroad Station and south of Brush Creek Ranger Station.

Amphibolites.—Thick dark medium-grained amphibolites occur together with the biotite schists in the Marble Creek Area (Loc. 364) and southeast of it. The main constituents are hornblende and plagioclase. Calcite lenticles ranging from a few cm. to several meters long are abundant. Biotite is a common dark constituent also in these layers and its distribution is rather irregular. These amphibolites most likely represent marly layers in the sedimentary strata. Dark amphibolites with abundant elongated epidote lenticles occur near the granite contact 1¼ miles south of Milsap Bar (Loc. 397). Some feldspar occurs in thin layers which may represent original bedding. A few large crystals of diopside can be detected in the epidote lenticles. Here higher content of quartz and alumina in the original sediment permitted the crystallization of abundant epidote, while in the amphibolites of the Marble Creek area calcite occurs in excess.

FIGURE 7.—HORNBLENDE-CHLORITE SCHIST. Large chlorite (chl II) flakes growing across the schistosity include hornblende (ho) needles. 300 meters south of Pulga, Loc. 41. 27×.

Older Intrusives

Serpentines and soapstones.—Serpentines and soapstones occur as lenses elongated parallel to the foliation among the metavolcanics and metasediments. Serpentines are dark greenish-gray to green and show a blocky weathering with shiny slicken sides. Main constituents in the serpentines are antigorite, chrysotile, and magnetite. Parallel orientation of minerals in streaks is common. Chrysotile seams in green serpentine south of Pulga may range up to 3 cm. in thickness. Every degree of serpentinization can be seen under the microscope. The remnants of the primary minerals are olivine, but shapes of augite can be detected in colorless amphiboles where serpentinization is incomplete, as is the case in a coarse-grained ultrabasic rock along the road from Brush Creek to Milsap Bar (Loc. 188). The thin section made from this rock shows the beginning of serpentinization of a peridotite (Fig. 8). The rock (No. 188) occurs together with serpentines

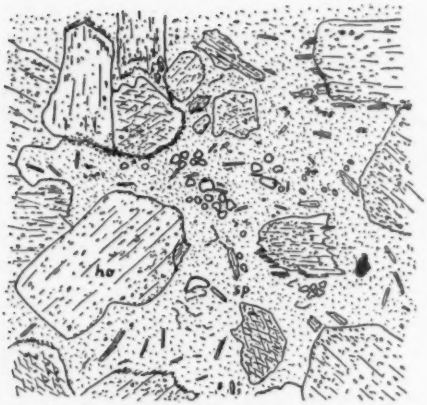

FIGURE 8.—SERPENTINIZATION OF PERIDOTITE
Remnants of olivine (ol) in serpentine (sp). Many of the hornblende (ho) show shape of augite and includes abundant magnetite (black). West of Milsap Bar, Loc. 188. 21×.

FIGURE 9.—SERPENTINIZATION OF TREMOLITE (tr)
Mouth of French Creek, Loc. 179. 23×.

and talc schists which probably are more complete alteration products of the same intrusive. The larger amphiboles which show outlines of augites include numerous small magnetites. The optical properties $\gamma = 1.631 \pm 0.001$, $\beta = 1.619 \pm 0.001$, $\alpha = 1.605 \pm 0.001$, $\gamma \wedge C = 17°$ indicate it to be a tremolite with 5 per cent actinolite. The remnants of olivines are embedded in a mass of antigorite with small tremolite needles. The areas occupied by these minerals are free of magnetite. The original rock consisted of about 40 per cent olivine and 60 per cent augite. The chemical analysis is shown in Table 1, No. 9.

Remnants of olivine were found also in fine-grained serpentine dikes 0.7 miles north of Pulga Railroad Station. These dikes occur in biotite schist, talc schist, and coarse-grained green serpentine. They consist of antigorite with some dust-like magnetite. Groups of small grains of olivine are embedded in antigorite. The border zones, cracks, and a few coarse-grained spots contain talc. These dikes are clearly younger than the main serpentine body in which they occur.

The border zones of large serpentine lenticles are usually converted to soapstones or to talc schist. The minerals in the soapstones are talc, magnesite, serpentine, magnetite, and chlorite. Addition of CO_2 is necessary in this reaction which can be expressed as follows:

$$\underset{\text{2 Serpentine}}{2H_4Mg_3Si_2O_9} + 3CO_2 \rightarrow \underset{\text{Talc}}{H_2Mg_3Si_4O_{12}} + \underset{\text{3 Magnesite}}{3MgCO_3} + 3H_2O$$

The talc schists consist of talc, chlorite, serpentine, and magnetite. In the contact zones with siliceous country rocks serpentine reacted with silica and was converted to talc according to the following equation:

$$H_4Mg_3Si_2O_9 + 2SiO_2 \rightarrow H_2Mg_3Si_4O_{12} + H_2O$$

Tremolite has been altered to serpentine near the mouth of French Creek (Loc. 179). Several rather thin layers of serpentine and talc schist are intercalated with amphibolites (probably basic tuffs in origin) and mica schists. Microscopic investigation of sp. 179 shows that this serpentine contains remnants of colorless tremolite needles whose refractive index is $\gamma = 1.627 \pm 0.001$. The cracks perpendicular to the prismatic cleavage are filled by a similar greenish, fibrous chrysotile as in the main part of the rock (Fig. 9.) Penninite with small 2V and some magnetite are additional constituents.

CaO must be removed in solution together with silica when chrysotile replaces tremolite. Turner (1947) has pointed out that serpentinization usually takes place without change in volume. MgO and SiO_2 must be removed

from olivine since water is added to the system. Turner points out the difficulty in explaining how and when this magnesia is removed since signs of magnesia metasomatism in the country rocks are usually lacking. In the area under consideration, serpentinization has affected not only olivines, but pyroxenes also are converted to serpentine through various stages. It was not MgO but CaO which was removed from the system. Tremolite, which usually replaces pyroxene in the low grade metamorphism, may be replaced by the equal volume of chrysotile according to the equation:

$$2Ca_2Mg_5Si_8O_{22}(OH)_2 + 5MgO + 8H_2O \rightarrow 5H_4Mg_3Si_2O_9 + 6SiO_2 + 4CaO$$

Magnesia, which is freed during the serpentinization of olivine, replaces lime in pyroxenes and causes their serpentinization. Lime and silica which must be removed may have caused further metasomatic alterations in the surrounding rocks. We may thus expect to find lime metasomatism rather than magnesia metasomatism in the region surrounding the ultrabasic rocks which are composed of olivine and calcium-bearing pyroxene, or only of various pyroxenes.

In sp. 188, pyroxenes were first converted to tremolite. The peridotite which consisted of olivine and augite was converted to serpentine-tremolite rock with remnants of olivine. The advancing serpentinization would convert the rest of the olivines and tremolite to serpentine.

A schist exceptionally rich in chlorite was found only in one locality in the area, Loc. 74, which is near the contact of a large serpentine body south of Pulga. Chlorite occurs in minor amounts in all schists in the area, but it is a result of low temperature metamorphism and alteration rather than a product of metasomatism. Talc is a common constituent of the serpentines, and many smaller lenticles are converted completely to talc-chlorite schist. No enrichment of chlorite occurs in the contact zones of these lenticles.

While doing a geologic mapping and field investigation for the mining company Suomen Mineral O.Y. in Paakkila in eastern Finland, the writer saw quite different contact zones around low metamorphic products of ultrabasic intrusions. There the ultrabasic bodies were converted to anthophyllite (asbestos), serpentine, and talc; and as a rule these rather small bodies were enveloped by a biotite layer of considerable thickness, indicating outward migration of MgO and FeO.

Talc schists.—Thin layers of fine-grained, talc-tremolite-chlorite schist, common everywhere in the area, are especially numerous near the serpentine bodies and may represent ultra basic dikes. The small needles of tremolite with $\gamma = 1.640 \pm 0.001$ and $\gamma \wedge C = 20°$ measured in sp. 286 are the main constituents, but talc is abundant enough to make the schist greasy to feel. Younger tremolite needles, growing across the lath-shaped chlorite and talc, indicate the addition of the lime and silica into these layers.

Tremolite rocks.—In addition to the tremolite-bearing rocks of sedimentary origin, there are layers and lenticles, which consist of solid tremolite. Tremolite occurs in long prisms oriented at random. These dikes and lenticles may have consisted of ultra basic rocks, pyroxenites in which there was a considerable amount of lime originally or, part of the lime may have been added metasomatically.

Metagabbros.—A large body of a fine- to medium-grained massive or slightly schistose dark amphibolite in the Big Bar Lookout area most likely represents a metamorphosed basic intrusion. The minerals are hornblende, plagioclase, quartz, and epidote. Sphene and magnetite are the accessories. Numerous narrow veinlets consisting of quartz and epidote, transect the rock irregularly; 70 to 80 per cent of the rock consists of green hornblende. The large crystals are bent showing post-crystallization deformation. Recrystallized needle-like hornblende transects the light minerals. Abundant epidote occurs as inclusions in plagioclase.

Altered trondhjemite. — An intrusive, light-brownish rock consisting mainly of plagioclase, quartz, and some micas occurs as a lenticular body in the Big Bend area. This body is surrounded by quartz porphyry in which euhedral quartz phenocrysts, ranging from one-half to 1 cm. in diameter and somewhat smaller albite phenocrysts, are abundant. The fine-grained soda-rhyolite joins them on the east and south side and dacitic metatuffs in the west. A gold

quartz vein occurs between the intrusive and tuffaceous rocks. The intrusive near this vein appears to be altered hydrothermally. Abundant sericite and epidote are included in large

FIGURE 10.—ALTERED TRONDHJEMITE
Inclusions of epidote and sericite in the blocky plagioclase (pla) in the altered trondhjemite No. 328. chl = chlorite, ep = epidote. Surcease Mine, Big Bend. 19×.

square plagioclase crystals (Fig. 10). Most of the quartz is interstitial, but some grains show euhedral forms against plagioclase. A part of the epidote is associated with chlorite, which contains some rutile needles and is clearly an alteration product after biotite. A few garnets occur as an accessory.

The contacts between quartz porphyry and the intrusive body are not exposed, but the field relations suggest that the quartz porphyry is a border facies of the same mass. The fine-grained extrusives of this area belong to soda-rhyolites. The chemical composition of the intrusive rock (Table 1, No. 7) appears to be similar to that of the trondhjemites. The identity becomes more obvious if the amounts of calcite, muscovite, and epidote introduced hydrothermally are subtracted from the analysis.

Older Dikes and Sills

Ultra basic dikes and sills.—Serpentine dikes 0.7 miles north of Pulga are described in connection with the main serpentine body.

A coarse-grained, greenish-gray dike transects a serpentine body near Yankee Hill junction about 5 miles south of Pulga bridge (Loc. 103). The minerals are antigorite with $\alpha = 1.576 \pm 0.001$, $\beta = 1.579 \pm 0.001$, $\gamma = 1.582 \pm 0.001$ and $-2V$ large; and tremolite with $\alpha = 1.613 \pm 0.001$, $\gamma = 1.638 \pm 0.001$ which values indicate about 11 weight per cent actinolite and 89 per cent tremolite. Antigorite aggregates in the hand specimen are dark green and many of them show external shapes of euhedral olivine. Aggregates showing external shapes of augite consist of coarse-grained tremolite while the main part of the rock consists of fine-grained tremolite. The rock was originally olivine-augite porphyry. About 30 per cent of the rock consists of antigorite (former olivine phenocrysts) and 70 per cent tremolite. Thus, the composition of this dike is near that of peridotite No. 188. Some magnetite and pyrrhotite occur as accessories. A hornblendite, which crops out on a road cut about 1¼ miles north from the Pulga bridge (Loc. 25), was probably originally a pyroxenite.

Gabbroic dikes and sills.—Metamorphosed gabbro dikes and sills were found in several localities 98, 184 (a fourth of a mile south of Coyote Gap) 227, 301, and along the Gravel Range road. Specimen No. 98 is a greenish-gray, medium-grained rock consisting of hornblende, antigorite, plagioclase and some quartz. Antigorite aggregates suggest the presence of olivines in this rock before metamorphism. A part of the hornblende occurs in long prisms with well developed prism faces (110) but many hornblende crystals show outlines of augite. Thus the original gabbros most likely consisted of augite, olivine, hornblende, and some plagioclase. A few small grains of magnetite occur as an assessory. Large colorless hornblende crystals with $\gamma \wedge C = 24°$ form the main part of the gabbroic dike No. 184. Abundant small needles of tremolite transect the large plagioclase crystals. Zoizite, occurring as a minor constituent, indicates a large amount of anorthite in the plagioclase prior to metamorphism. This dike and a similar dike along the Gravel Range Road (Loc. 247) occur in serpentine.

Porphyritic dikes near Las Plumas (Loc. 301), consist of hornblende, calcite, epidote, quartz, plagioclase, and chlorite. Hornblende occurs in short euhedral prisms of medium size

(about 1 x 2 mm.) and form about 50 per cent of the rock. Calcite is abundant in the groundmass (about 50 per cent); quartz and plagioclase (An_{30}) occur as separate grains or form together a granophyric groundmass. Chilled border zones are fine-grained.

The sill rock No. 227, is a coarse- to medium-grained uralite porphyry. Uralite crystals range from 2 to 4 mm. in diameter and are embedded in a medium-grained ophitic groundmass. The interspaces between euhedral plagioclase laths are filled by hornblende, epidote, and chlorite. Epidote occurs also as larger grains and groups of grains.

Uralite porphyry No. 175 is another basic sill or dike which is later than the extrusives in the same area. It is chemically (Anal. 1) closer to the gabbro, No. 270, (Anal. 14) than to the metabasalt, No. 156, (Anal. 2). Rock No. 270 occurs as a large inclusion in the granodiorite and represents a basic member of the granodiorite series. Uralite porphyry, No. 175, is probably one of the basic forerunners which are also common elsewhere in the Sierra Nevada as described by several authors (Mayo, 1935; Erwin, 1934). The following minerals occur in No. 175: plagioclase, hornblende, chlorite, apatite, and ilmenite-magnetite.

Part of the plagioclase (An_{32-33}) occurs in large phenocrysts, whose centers contain some sericite, epidote, and magnetite. Many uralites still show external shapes of augites. Their optical properties are: γ (Blue-green) = 1.685 ± 0.001, α (pale green) = 1.646 ± 0.001, $\gamma \wedge C = 21°$.

Metadiabase, extending from the railroad cut 1 mile north of Intake Dam westward to the road going from the Big Bend road to Bardees Point (Loc. 343, 344, and 352), is a medium-grained slightly schistose dark-gray rock. The minerals are hornblende, epidote, plagioclase (An_{33}), some quartz, chlorite, and magnetite. Sphene and ilmenite are altered to leucoxene and magnetite is altered partly to hematite. Porphyritic texture is recognizable even in the coarse-grained center of the sill (No. 352). Uralite occurs as larger grains, and square blocky plagioclase crystals of medium size form the groundmass. Newly crystallized epidote, ranging from 0.1 to 2 mm. in diameter, may be found anywhere regardless of the boundary lines of earlier minerals. Many of the plagioclase crystals are completely altered to epidote. Aggregates, consisting of epidote and quartz and showing the external shape and cleavage of augite, were found in Loc. 344. The original cleavage cracks were filled by ilmenite and magnetite. Part of the hornblende occurs as large crystals and part as small needles.

The finer-grained dark border facies north of 344 (Loc. 346 as well as 343) shows a diabase structure. Plagioclase (An_{28}) occurs in long laths and hornblende fills the interspaces. Epidote occurs in stringers and as inclusions in the plagioclase and hornblende.

Dacite porphyry and quartz porphyry.— A white-weathering, fine- to medium-grained sill rock near the mouth of Berry Creek (Loc. 140) consists of plagioclase (An_{29}), quartz, muscovite, epidote, and biotite. The fringed lath-shaped plagioclase crystals show a trachitoid texture. Small quartz and plagioclase grains are interstitial and show in many places a granophyric intergrowth. Some muscovite follows small cracks and may have been introduced hydrothermally. The amount of plagioclase is about 50 per cent, which is larger than that of the effusives in the area. It is mineralogically close to the trondhjemite, No. 328, and together with it represents connecting members between the older effusives and younger intrusives. Dikes and small sills of a similar composition occur also along the railroad east of this locality (Loc. 176) and near Poe Railroad Station.

The medium-grained, gray dike rocks, Nos. 383 and 393, in the Bloomer Hill area consist of plagioclase, pale-green to colorless amphibole, quartz, epidote, and some chlorite. Euhedral and square grains of plagioclase are altered to a fine-grained mass of epidote and zoisite forming about 50 per cent of the rock. In No. 393, some plagioclase occurs in larger phenocrysts and the groundmass is granophyric. Plagioclase in it is unaltered, suggesting more albitic composition. Rock No. 383 is equigranular. Both of these hypabyssal rocks and the dacites in which they occur have a similar composition.

The same relation was found between metarhyolite, No. 288, and a quartz porphyry, No.

289, near a spring 2½ miles south of Bear Ranch Hill Lookout. The mineral composition in both is the same: quartz, feldspars, biotite, muscovite, and epidote. Quartz and plagioclase phenocrysts are strongly deformed and granulated. Muscovite and epidote are secondary after feldspars. Biotite is recrystallized and concentrated in stringers in the fine-grained groundmass.

Granodiorite and Related Intrusives

Introduction.—Like the larger Sierra Nevada batholith south of the area under consideration, the small batholiths of this region are younger than the metavolcanics and metasediments. The prevailing intrusive rock type in the present area is a granodiorite, but more acidic and more basic differentiates also occur.

Gabbro.—Gabbros and hornblendites are found as small intrusions in the older rocks near the contacts of granodiorite plutons (Loc. 377) and in larger quantities in the northern part of the Bidwell Bar Quadrangle, in the zone from Bucks Mountain to the Feather River Highway between Storrie and Tobin. No. 14 in Table 1 shows the chemical analysis of a basic hornblende gabbro south of Rock Creek (Loc. 270). The main dark mineral in this rock is blue-green hornblende with the following optical properties: highest γ measured = 1.685 ± 0.001 lowest α measured = 1.630 ± 0.001, $-2V = 74°$, in the thin section $\gamma - \alpha = 0.036$, $\gamma \wedge C = 17°$. The variation of the refractive indices indicates some variation in the chemical composition. This variation also is seen in the spotted color of the hornblende crystals. Plagioclase (An_{36}) contains numerous inclusions of epidote and muscovite. Chlorite occurs as an alteration product after biotite. Magnetite is an accessory. Also in gabbro No. 359 (200 m. north of Storrie, Loc. 359 on Pl. 1), the hornblende is bluish green. The centers contain abundant small inclusions of scaly ilmenite oriented parallel to the cleavage or some other crystallographic plane. The border zones are free of these inclusions. Probably the centers originally consisted of augite which was altered to hornblende during the latest period of crystallization. The refractive index α = 1.646 ± 0.001 suggests that this mineral has about the same chemical composition as the hornblende with higher indices in No. 270. Plagioclase (An_{36-44}) and some quartz are the light minerals and constitute about 20 per cent of the rock. Biotite and apatite are the minor constituents.

Diopside was found in only two specimens investigated: in gabbro 377 and diorite 358. In both cases, it is surrounded by amphiboles. Gabbro 377 is a medium-grained, gray rock in which dark and light minerals can easily be recognized with the naked eye. The minerals determined under the microscope are plagioclase (An_{50}), diopside, amphiboles, and epidote. Diopside is partly altered to green hornblende and partly to colorless cummingtonite. Fibrous bluish-green hornblende occurs along the border zones of larger amphiboles and as inclusions in altered plagioclase. The square centers of the plagioclase crystals consist of epidote indicating that they were rich in anorthite. The rims are unaltered and consist of An_{50}. The alterations of diopside and anorthite are probably deuteric and caused by the lowering of the temperature during the latest phase of crystallization. The clear rims of plagioclase most likely crystallized during the same period.

Diorite.—Diorites, like the gabbros, occur in small lenticles along the border zones (Milsap Bar and north of Storrie, Fig. 11) or as transitional members between gabbros and granodiorites (near Grizzly Creek and Storrie). The diorites are medium- to coarse-grained gray rocks in which plagioclase, quartz, hornblende, and biotite can easily be recognized with the naked eye. A photomicrograph of diorite No. 118 is shown in figure 1 of Plate 4. Determination by the universal stage shows that the albite, Ala B, Carlsbad, pericline, and complex twinnings are common in the plagioclase. Composition varies from An_{34} to An_{44}. Hornblende is of a common green variety with $2V = 66°$, $\gamma \wedge C = 23°$, and α = 1.661 ± 0.001 measured in No. 118. Biotite shows γ = 1.660. Magnetite, apatite, sericite, epidote, and allanite occur in minor amounts. Zeolite occurs as an alteration product in one zoned plagioclase. Some unaltered diopside surrounded by hornblende was found in diorite 358. Chemical analysis of diorite 118 is No. 16 in Table 1.

Granodiorite.—Granodiorite is the most com-

mon rock type among the younger intrusives. The Merrimac pluton, situated in the center of the mapped area, consists almost exclusively of granodiorite. Only a small portion of the center

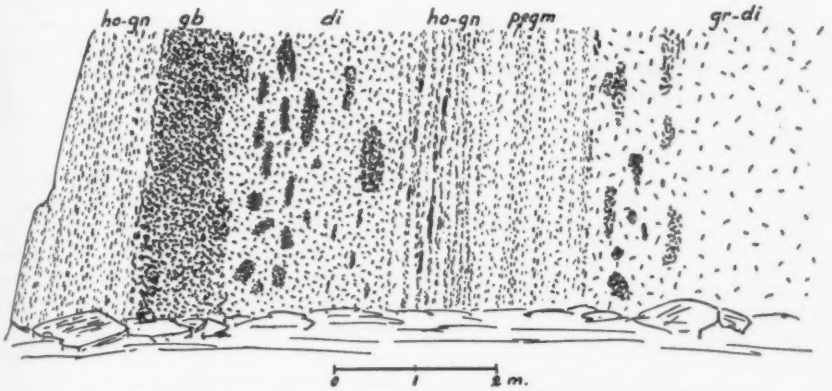

FIGURE 11.—CONTACT ZONE BETWEEN GABBRO (gb) AND GRANODIORITE (gr-di) North of Storrie. di = diorite, ho-gn = hornblende-gneiss, pegm = pegmatite vein.

shows a greater acidity with a granitic-trondhjemitic composition. The mapped portion of the northern pluton also consists mostly of granodiorite mineralogically similar to that in the Merrimac pluton. The mapped part of the large pluton in the southeastern corner of the map consists mainly of medium-grained, light-gray granite, and the more basic varieties occur only locally near the contacts with amphibolites like the diorite east of Milsap Bar.

The granodiorites are lighter in color and of somewhat smaller grain size than the diorites. Hornblende and biotite are similar to the corresponding minerals in the diorites. Epidote occurs together with dark minerals. In many localities (north of Storrie, Loc. 123 and south of Wild Cat) it is found also as an abundant alteration product in the plagioclase. Zoned plagioclase (An_{30-28}) and the dark minerals are euhedral, and quartz is interstitial. Microcline is a common additional constituent (Fig. 12). Magnetite, sphene, apatite, and allanite occur as accessories. Muscovite may appear together with epidote as an alteration product in plagioclase. Chloritization of biotite is common but usually minor as also is myrmekitic intergrowth of quartz and plagioclase.

Chemical analysis of granodiorites Nos. 119 and 121 north of Grizzly Creek are shown in Table 1, Nos. 17 and 18.

The texture of the granodiorite, No. 119 (Fig. 12), as well as that of diorite, No. 118

FIGURE 12.—GRANODIORITE

Minerals: qu = quartz, pla = plagioclase, mi = microcline, bi = biotite, ho = hornblende, ep = epidote, black = magnetite. ½ mile north of the mouth of Grizzly Creek. Loc. 119. 32×.

(Pl. 4, fig. 1), differs considerably from that of the Merrimac granodiorite and Bald Rock granite (Pl. 4, fig. 2). In the two latter rocks, the plagioclase is euhedral and zoned. The interstitial quartz and microcline occur in large grains including the plagioclase. In contrast to this, the plagioclase in the diorite, No. 118, and

in the granodiorites of the same area is zoned only slightly or not at all. It occurs in long subhedral crystals with curved boundary lines. Quartz and microcline appear in small grains

TABLE 3.—GRANITE OPPOSITE ENTERPRISE, NORTH SIDE OF THE SOUTH FORK OF THE FEATHER RIVER

(Turner 1894)

SiO_2	70.36
TiO_2	0.20
Al_2O_3	15.47
Fe_2O_3	0.98
FeO	1.17
MnO	trace
MgO	0.87
CaO	3.18
BaO	0.06
SrO	trace
Na_2O	4.91
K_2O	1.71
LiO_2	trace
P_2O_5	0.11
H_2O^+	1.00
H_2O^-	0.06
	100.08

between the plagioclase crystals. The femic minerals hornblende and biotite, in the Merrimac granodiorite, and Bald Rock granite show more euhedral forms than the same minerals in the diorite and granodiorite of the Grizzly Creek area. Diorite, No. 118, and granodiorite, No. 119, occur as contact facies near the brecciated zone at the mouth of Grizzly Creek. There is a complete gradation from the brecciated amphibolite to the coarse gabbro, as well as to diorite, No. 118, and granodiorite, No. 119, which thus represent two steps in the digestion of the amphibolite blocks by the granodioritic magma, No. 121.

Granite-trondhjemite.—The granites and trondhjemites are lighter in color than the granodiorites because of a smaller amount of dark minerals. Biotite is usually the only dark constituent. Quartz and zoned partly euhedral plagioclase (An_{20-26}) are the main light constituents (Pl. 4, fig. 2). The amount of microcline varies but always is much less than that of plagioclase. Muscovite, epidote, and magnetite occur in minor amounts. Epidote may abound near the contact zone as in Loc. 260. Myrmekitic intergrowth of quartz and plagioclase is common.

Chemical analysis of a specimen from the Little Bald Rock is shown in Table 1, No. 19. This as well as an old analysis of a granite specimen from the southern part of the same pluton (Table 3) shows a lower percentage of potash than is present in normal granites. The relation $Na_2O:K_2O$ is indeed the same as in the normal trondhjemites. In the trondhjemite from Little Bald Rock, 84 per cent of the total feldspars is plagioclase (An_{17}). The same relation was observed in the acidic coarse center of the Merrimac pluton.

The contact phase of the Bald Rock pluton consists of a gneissoid granite in which relicts of older rocks, meta-volcanics, can be easily identified (Pl. 3, fig. 4).

Pegmatite.—Pegmatite veins and dikes are common especially near the contacts of the granitic and granodioritic intrusions. They cut the country rocks as well as the intrusives. In the intrusives they usually follow foliation or joints ranging mostly from 2 to 50 cm. in thickness. Analysis No. 20 shows the chemical and mineralogical composition of a pegmatite dike occurring in diorite on the north side of the mouth of Grizzly Creek (Loc. 116). This pegmatite is very light gray. It consists of quartz, plagioclase (An_{13}) and orthoclase in about equal amounts. A few muscovite and biotite flakes occur as additional constituents.

Younger Dikes

Introduction.—The younger dike rocks may be divided petrographically into: (1) the porphyritic dikes which range from gabbros to diorites and granodiorites or plagioclase porphyries in composition; and (2) the nonporphyritic dikes, consisting of albitic plagioclase and hornblende in varying amounts, or of albite and biotite. The latter group is characterized by abundance of albitic plagioclase and absence of quartz and microcline. Hornblende or biotite are the dark minerals. Oligoclase gabbro, albite diorite, and biotite albitite can be distinguished.

Porphyritic gabbro and diorite dikes.—This group is represented by porphyritic gabbros and diorites of normal composition and by plagioclase porphyries.

Porphyritic gabbro dike No. 360 from the Marble Creek gorge contains large plagioclase phenocrysts which are altered to epidote with some sericite. Smaller phenocrysts consist of unaltered plagioclase (An_{35}) and hornblende needles. Both are euhedral. The minerals in the ground mass are small hornblende needles, quartz, plagioclase, epidote, sericite, and magnetite.

In the plagioclase porphyry dikes also (Locs. 416 1 mile E. of Big Bar Hill and 296 SW. of Walker Plain) epidote is an abundant alteration product especially in the phenocrysts. The ground mass is fine- to medium-grained and consists of plagioclase, quartz, biotite, epidote, hornblende, and some chlorite and magnetite. The mineralogical composition is like that of dacite porphyries.

Dioritic dikes near Milsap Bar and south of Wild Cat contain hornblende as phenocrysts, with epidote as a common secondary mineral. The fine-grained, granoblastic ground mass consists of hornblende, plagioclase, quartz, and epidote. Hornblende is oriented at random in the center of the dike but parallel to the walls along the border zones.

Trondhjemite porphyry.—The acidic porphyritic dike rock in the gorge of Stony Creek (Loc. 182, 368) is a fine-grained, gray rock in which plagioclase (An_{20}) occurs in abundant euhedral phenocrysts. The groundmass is fine-grained and consists of quartz, plagioclase, hornblende, epidote, and magnetite (Pl. 3, fig. 5). The mineral composition is similar to that of plagioclase porphyries but with more quartz and less dark minerals. The composition is like that of the hornblende trondhjemites. Thus we seem to have also among the porphyritic dikes a differentiation series which is characterized by a small amount of potash.

Rock No. 368 is a porphyritic trondhjemite. The ground mass shows an abundant granophyric intergrowth of quartz and plagioclase. No potash feldspar occurs in this rock. The dark minerals are biotite, chlorite, epidote and magnetite. A similar rock occurs as veins in talc-tremolite rocks 5 miles south of Pulga (Loc. 80). Large chlorite flakes are surrounded by hornblende needles. Granophyric intergrowths of quartz and plagioclase occur locally.

Oligoclase gabbro—albitite series.—The oligoclase gabbro dikes (Nos. 32, 44, 403, 405) are medium- to fine-grained dark rocks in which the rather short hornblende prisms are oriented

FIGURE 13.—LAYERS IN GABBRO DIKE
Contact of a medium and fine-grained gabbro layer in a basic dike 2 miles NW. of Merrimac. Minerals: plagioclase (pla), hornblende (ho) and epidote (ep). 20×.

at random. Many of the dikes are medium-grained near the center and fine-grained near the borders. Hornblende is oriented parallel to the walls in the fine-grained border zones. Some dikes show alternating layers of medium-grained and fine-grained gabbro (Fig. 13); and, in others the medium-grained lenticles are included in a fine-grained rock showing a flow structure. These dikes are abundant in the Merrimac granodiorite area and occur usually parallel to the joints of the plutonic rock (Fig. 14). The border zones of wider dikes include long sheets of wall rock, granodiorite, which shows no alteration even next to the contact. The contacts between the granodiorite and the dikes usually follow straight lines. Plagioclase An_{26} and hornblende are present in equal amounts. Some apatite and magnetite occur as accessories.

The mode of occurrence of the albite diorite dikes (No. 42) is similar to that of oligoclase gabbros. The color is somewhat lighter because of a smaller amount of hornblende. The same type of dikes cut the pre-granitic rocks south

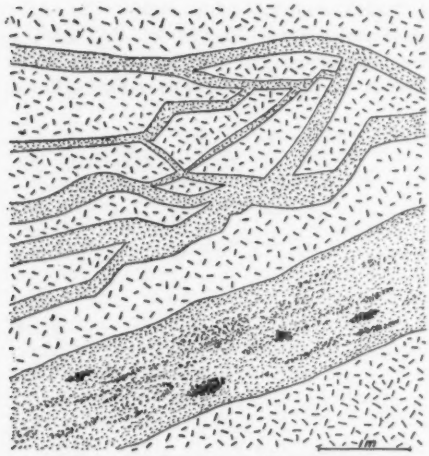

FIGURE 14.—OLIGOCLASE GABBRO DIKES IN GRANODIORITE
3 miles northwest of Merrimac.

The most acidic members of the series are fine-grained, gray rocks in which albite (An_7), occurring in long laths with scarce perthite, forms from 70 to 80 per cent of the rock. Biotite is the dark constituent (Pl. 3, fig. 6). This type of dike was found in the Marble Creek gorge (Loc. 307) cutting the pre-granitic rocks.

These dike rocks, oligoclase gabbro, albite diorite, and biotite albitite, are closely related petrologically and seem to form a differentiation series which crystallized out of a magma poor in silica, calcium, and potash.

Walker Plain Basalt

The youngest rock in the area under consideration is a basalt which lies above the granodiorite and the older rocks on both sides of the old Quincy-Oroville road extending south

FIGURE 15.—MAIN OXIDES IN THE METAVOLCANIC AND INTRUSIVE SERIES
Solid lines = metavolcanic rocks; broken lines = intrusive rocks.

of Pulga and in the Marble Creek area concordantly and discordantly. Plagioclase varies from An_5 to An_{15} and makes up from 70 to 80 per cent of the rock. Hornblende, in long slender needles oriented at random, is the only dark constituent. Accessories are the same as in the oligoclase gabbro.

from Four Trees junction. The abundant phenocrysts in this rock consist of plagioclase and augite. Magnetite, partly altered to hematite, is a minor constituent. The groundmass is glassy. Plagioclase occurs in long laths oriented at random. They range from An_{60-80} in the centers to An_{48-55} at the borders.

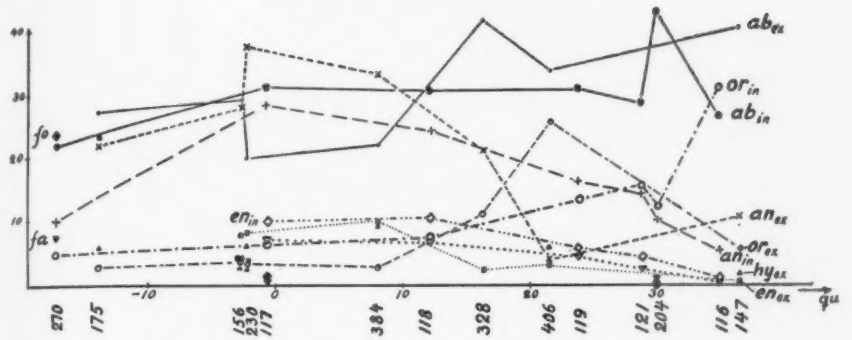

FIGURE 16.—NORMS OF THE METAVOLCANIC AND INTRUSIVE SERIES

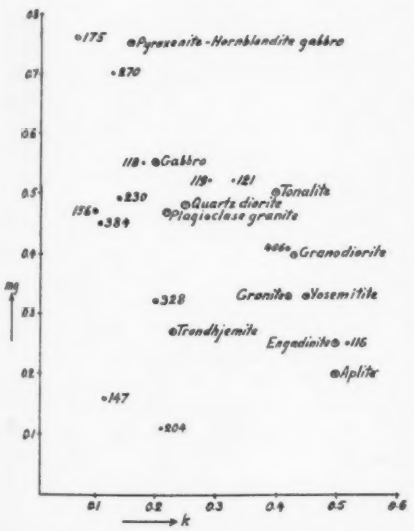

FIGURE 17.—k and mg OF THE METAVOLCANIC AND INTRUSIVE ROCKS

Niggli's normal magma types are indicated for comparison (circles).

COMPARISON OF THE CHEMICAL COMPOSITION OF THE METAVOLCANICS AND INTRUSIVES

The main oxides of the metamorphosed extrusives and intrusives of the Sierra Nevada series are plotted in Figure 15. There is a close similarity between these two magma series. In both, lime dominates over ferrous oxide and that over magnesia except in the most basic member of each series, in which case the amount of lime is larger than that of iron oxide but smaller than that of magnesia. Another exception in the extrusive series is No. 406 which was found only in one locality and considered to be a product of biotite enrichment of a soda-rich rhyolite. The amounts of K_2O and Na_2O vary in the acidic members of both series. Mineralogically this is visible in the varying amounts of potash feldspar. Some acidic extrusives like Nos. 147, 190, 210, and 330 contain no potash feldspar, in some like No. 221 the amount of potash feldspar is about half of the amount of plagioclase. Potash feldspar is always present in the intrusives but as a rule in an amount only about a fifth of that of plagioclase. In the pegmatites these two feldspars occur in equal amounts. Thus both series are rich in soda. Predominance of the soda over the potash is especially striking among the dike rocks of both series. The only well preserved acidic intrusive stock belonging to the extrusive series is the trondhjemite porphyry No. 328 in which there is no potash feldspar. It corresponds thus to the soda-rhyolite No. 147 among the extrusives and may represent the throat of a volcano, from which the surrounding soda-rhyolites were extruded. On the other hand, hornblende-albite rocks are common among those dikes, which are younger than the Sierra Nevada intrusive series.

In the norm (Fig. 16), albite strongly dominates over orthoclase and the Niggli value k is fairly low in both series (Fig. 17). Pegmatite 116 is the only exception in the intrusive series. The calc-alkali index for both series is between 60 and 65 (Fig. 15), indicating that both belong to the calc-alkalic magma group.

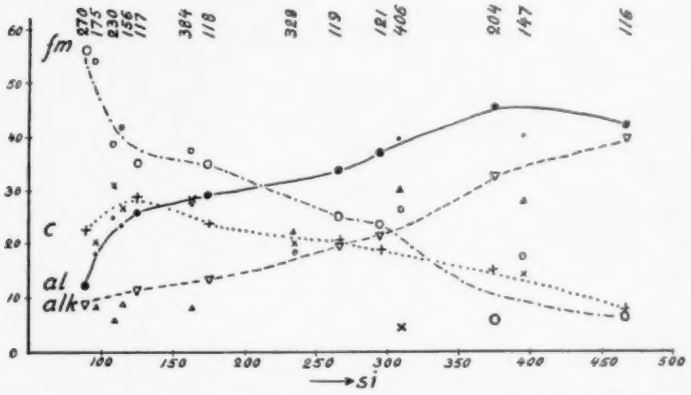

FIGURE 18.—NIGGLI VALUES OF THE METAVOLCANIC AND INTRUSIVE SERIES

The Niggli values al, fm, c, and alk are plotted in Figure 18. Curves are drawn only for the intrusive series but the diagram shows that the corresponding values for the extrusives fall near the same curves.

Thus both magmatic series show the same chemical characteristics and seem to be closely related. Trondhjemitic differentiates occur together with the granitic ones among the most acidic members.

Lindgren has pointed out that the magmatic activity in western United States was moving towards the east during the Mesozoic and at the same time there was a clear tendency of the magmas to become more alkaline (Lindgren 1933). The same tendency towards the alkaline series is seen in the present area when the chemical composition of the older effusive series, younger intrusives and the youngest dike rocks are compared:

The effusive rocks are somewhat richer in silica and poorer in alkalies than the corresponding intrusive rocks (Fig. 15). The youngest series, the dike rocks, is still richer in alkalies especially in Na_2O. Most dikes consist only of two minerals, hornblende and sodic plagioclase in varying amounts. The most leucocratic variety among these dike rocks is biotite-albitite (No. 307).

STRUCTURES

Structures of the Metamorphic Rocks

Introduction.—The metavolcanics, intercalated sedimentary rocks, and older intrusives are folded and deformed together and thus show similar structures. The strikes and dips of the bedding and foliation are indicated on the map by structure marks. As a rule the basaltic flows of the same area have been less resistant to the deforming movements and metamorphic agencies than the rhyolites. This may be partly due to the fact that the primary mineral assemblage of rhyolites was stable in those physical-chemical conditions under which metamorphism and contemporaneous deformation took place whereas an adjustment was needed to bring the mineral assemblages of dacites and basalts into equilibrium. Furthermore, the rocks rich in hornblende develop a schistosity more readily than those rich in quartz and feldspar.

Bedding.—Most of the sedimentary rocks show a distinct bedding. It is visible because of the alternating layers of quartzites, mica schists, hornblende schists, and limestones. In addition, the amount of dark and light minerals in separate layers of the schists and quartzites varies. The intense folding and transecting cleavage tends to obscure the bedding in the fine-grained chlorite-epidote schists near the mouth of the West Branch of the Feather River and in biotite schist of the Marble Creek area (Pl. 4, fig. 4). It can be detected in the fresh surface and in the thin sections, however, because of the variation in color and minerals, In the localities near Pulga and Milsap Bar, the bedding dips steeply. Most of the beds are rather thin. The quartzitic biotite schists are thinly laminated within each bed.

Hornblende-epidote schists of volcanic origin occur as thin layers in the sedimentary schists. Because of the discontinuity of the outcrops and lenticular structure of the sedimentary and volcanic formation, one cannot follow individual beds long distances. The lenticular structure is well represented on the map by the occurrences of limestones, quartzites, and metavolcanics.

Primary structures of the metavolcanics.—Pyroclastic structures are common in the metavolcanics. The dacite agglomerate from the northern slope of Bloomer Hill (Loc. 387, Pl. 2, fig. 1) is described under metadacites.

Volcanic breccia of basaltic composition is well exposed between the mouths of Berry Creek and French Creek (Loc. 163). Angular fragments of porphyritic basalt are cemented together by fine-grained material of the same composition. Square blocky plagioclase occurs as abundant phenocrysts in the fragments. Strongly deformed bombs of a dacite agglomerate east of Deer Park are shown in figure 3 of plate 2. There are fragments of a coarse to medium-grained gabbro among the fine-grained lava bombs in the undeformed center of this occurrence.

The agglomerates occur as lenticular bodies in metamorphosed tuffs and lavas or between lava flows and older sedimentary rocks. Some tuffaceous layers show a distinct layering like those near Pulga (Locs. 31, 59). In most cases, however, they are rather homogeneous, fine-grained, and schistose without any primary structure visible megascopically.

Folds and axis of folding.—Only in a few localities are folds so small that the folding axis can be determined directly in the field. Along the railroad east of Big Bend, the bedded mica-schists, probably originally rhyolite tuff, (Loc. 158) show rounded folds whose axis plunges 40° W. Folds whose axes plunge 30° N. W. can be seen in a light-gray amphibolite and in mica schist 2.2 miles south of Pulga Railroad station. Limestone beds near the bridge going over Marble Creek are folded together with amphibolite. The axis of a drag fold appearing on the northwest side of the bridge plunges 40° north.

Small folds can be detected in thin sections made from the schists near the mouth of the West Branch of the Feather River and from the Marble Creek area (Pl. 4, fig. 4). Foliation is the only structure which is readily observable in the outcrops. Bedding planes appear to be intensely folded and obscured by the shearing parallel to the cleavage, which intersects the bedding.

Lineation.—Lineation is not common in the metamorphic rocks of this area. It is marked on the map in those localities where it could be measured. Its direction may parallel the axis of the folding (*b*) or *a* axis which then appears to be the axis of cross folding, or there is no visible relation between it and other structural elements.

Where individual small folds can be seen in the field, the lineation was found to parallel the axis of folding, as in several localities along the North Fork of the Feather River, such as Las Plumas area, the eastern Big Bend area (Locs. 158, 314), several localities between Poe and Pulga Railroad station, and Loc. 109 south of Grizzly Creek and in the area around Big Bar Hill. Similar structure occurs in many localities east of Merrimac, along the Sky High Road, Loc. 133, Davidson Mine, and lower part of Marble Creek area.

In many localities, lineation occurs either parallel to the *a* axis of the tectonite or parallel to a second folding axis, which then usually coincides with the direction of *a* in a large fold. This type of lineation is common around Deer Park and in an area extending from there eastward (Loc. 134); also in some localities northwest of Milsap Bar, mouth of Berry Creek, in the Big Bend Mine, and at the Intake Dam. No relation to other structure was found in locality 81. Directions of the lineation are rather irregular in many localities such as Loc. 77.

Intensity of the deformation is irregular. Harder layers like volcanic flows and dikes, seem to have been very resistant to deforming movements, while tuffs and sediments were readily deformed. The contact zones between the harder and less resistant layers are strongly deformed because the slippage must have been greater along these zones than elsewhere. This can be seen in the agglomerate lenticles east of Deer Park. The border zones and the thinner agglomeratic beds intercalated with the tuffaceous layers show a strong deformation. The pebbles are drawn out into long spindles,

and their length averages about 10 times the diameter of the cross section (Pl. 2, fig. 3). The central parts of the lenticles are undeformed. The fragments are rounded or square with rounded corners. Similarly the amygdules in the metavolcanics may be round or oval because of lack of deformation in the resistant layers and deformation along the zones of weakness.

The pebbles in the conglomerates near Las Plumas are flattened and drawn out into lenticles, whose longest axis parallels the axis of the folds and ranges from 4 to 7 times the length of the shortest axis. The axis of medium size parallels the *a* axis of the folds and is about twice as long as that of the shortest one.

The irregularity in direction of lineation is most likely due to the inhomogeneity of the folded strata. The structure is highly lenticular; the harder compact lenticles of various sizes are embedded in softer and less resistant material. When the folding took place, the hard lenticles and layers yielded very little or not at all to the deforming movements. Therefore, the slippage along their borders must have been many times greater than in the homogeneous less resistant strata. Moreover, the direction of the slippage may easily vary in this type of inhomogeneous material resulting in the irregularity in direction of lineation.

Foliation.—Foliation may parallel the bedding or intersect it at various angles. Where bedding is pronounced, the angle between the bedding and foliation varies from 0° to 15°. In the fine-grained schists near Las Plumas and in the Marble Creek area, it parallels the axial plane of small folds (Pl. 4, fig. 4).

In the fine-grained hornblende-chlorite schists, hornblende needles parallel the foliation, but many of the large chlorite flakes grow across it and include small hornblende needles (Fig. 7). These chlorite flakes are younger than the finer-grained sheared flakes oriented parallel to the foliation (chl. II in Fig. 7). No structure can be observed to parallel these younger ones in the field. Most of these late chlorites occur in small lenticles together with coarse-grained quartz, which like chlorite is later than the main part of the rock.

Structures of the Intrusive Rocks

Platy and linear structure.—Foliation and lineation of the intrusive rocks are marked on the map where they could be measured in the field. As a rule the centers of the larger plutons are structureless, but the border zones show a distinct foliation and in some localities also lineation. The foliation follows the trend of the contact, is most distinct next to the contact, but becomes obscure towards the center. It dips 50°–80° away from the contact. Lineation is well developed only at a few localities but it may be pronounced even where the foliation cannot be measured. Both these structures become visible because of the orientation of biotite and hornblende.

Jointing.—The joint systems are marked on the map (Pl. 1) with long thin lines. There are usually 3 joint systems corresponding to a system of cross joints and two longitudinal joint systems, but in many localities two to three additional systems can be measured.

As a rule joints are well developed and followed by fine- to medium-grained diorite dikes, pegmatites, and quartz veins. In those localities where the foliation can be seen, one of the joints parallels it. Another well developed system is the cross jointing, which makes an angle from 70° to 90° with the plane of foliation. It dips 30°–90° along the contacts. In the French Creek area (Loc. 312) where the coarse-grained, light-gray granite-trondhjemite shows well developed nearly horizontal layering a cross joint strikes N. 25° E. and dips 85° northwest. One of the longitudinal joints parallels the layering and another is perpendicular to it. The structure here is typical of that on the top of domes. However, foliation in the granodiorite about half a mile east of this locality dips 70°–80° north northeast.

Inclusions.—Dark, basic inclusions are common in the diorites and granodiorites. Near the contacts they are elongated parallel to the lineation, but in the central part they appear to be rounded. There are three types of dark inclusions: (1) fragments of the wall rocks, xenoliths, (2) large inclusions representing basic members of the magmatic series, autoliths, (3) dark inclusions of uncertain origin.

(1) Fragments of the wall rocks are common

near the contacts. They consist of hornblende-biotite schist or amphibolite. A chemical analysis (No. 15) was made of an inclusion of this type. It is included in diorite 118 (Analysis No. 16). A comparison of the two analyses shows that there is more calcium, iron, and aluminum and less silica in the inclusion than in the surrounding intrusives. Comparison of analysis of a metabasalt, No. 2, and that of the inclusion, No. 15, shows a great similarity. There is a slight addition of SiO_2, Al_2O_3, and K_2O in the inclusions due to the partial granitization.

The diameter of the inclusions of the wall rock varies from a few inches to 2 or 3 feet. Even larger inclusions occur just next to the contacts and they are angular or drawn-out fragments of the wall. Many of them still show a bedding or schistosity similar to that present in the wall, but these structures usually are no longer parallel to the corresponding structures in the wall. This is the case near the Davidson mine along the Sky High Road, where the large amphibolite blocks included in granite near the contact clearly have been moved (Loc. 135). The platy structure in the granite makes a little swirl at the same locality, and when the amphibolite block was rotated by the swirling magma, the long dimension of the block coincided with the platy structure of the granite.

There is no reaction zone around the inclusions. Thus recrystallization, where it occurred, has been complete, resulting in a homogeneous xenolith, and the equilibrium between the inclusions and surrounding intrusive has been attained. This is easily understood if we bear in mind that the inclusions are rather small and that in the walls, the zones next to the contacts are composed of the minerals that occur in the intrusive. The inclusions as well as the walls next to the contact were soaked in granitic solutions and transformed into rocks in which the minerals are the same as in the intrusive, but which are chemically more basic because of larger amounts of femic minerals originally present. Thus only the facies which occurs next to the contact in the wall is represented in inclusions, while successive zones of contact metamorphism and metasomatic granitization can be seen in the walls outside of the immediate contact line.

(2) Autoliths in the diorite consist of hornblendite or gabbro, and in granodiorite of gabbro and diorite. Also, these inclusions vary in size: the diameter of smaller ones from an inch to 2 feet, while the larger ones may range up to 40 feet long. Analysis No. 14 represents a large gabbro inclusion in granodiorite near the mouth of Rock Creek. There is considerably more magnesium and less aluminum and silica in it than in inclusion No. 15.

(3) The third group is made up of inclusions whose origin cannot be readily determined. Pabst (1928) has made an extensive study of this kind of inclusions in the main Sierra Nevada batholith. In the present area, the autoliths of this type are common. They are perfectly rounded, 2–5 inches in diameter, and consist of dark-gray, fine- to medium-grained gabbroic or dioritic rock. The minerals are the same as in the enclosing granodiorite. Many show a few larger hornblende and biotite crystals which give a porphyritic appearance to the inclusion. The texture is hypidiomorphic as in the granodiorite. The only difference mineralogically seems to be a greater amount of hornblende and lesser amount of quartz and potash feldspar in the autolith.

In the northern part of the Merrimac pluton, many rounded autoliths show a texture and mineral assemblage similar to that of the younger diorite dikes. Hornblende occurs in long needles oriented at random, and part of the feldspar tends to form larger crystals. There seems to be a genetic relationship between the two. The dikes were emplaced shortly after the solidification of the granodiorite and represent the basic members of the diaschistic dikes. This kind of splitting of the magma into basic and acidic phases obviously does not take place only after the emplacement of the main intrusion, but the two phases can exist also prior to the emplacement of the granodiorite and the autoliths may represent an earlier phase of the basic offshoots.

Their porphyritic texture suggests that they were crystallized under the same condition as the dike rocks, then later on fragments of this basic phase were brought up by rising magma. At this point it may be well to point out again

the great petrographic similarity of the older effusive and younger intrusive series and remind the reader also that members connecting these two series occur. The dioritic dikes and

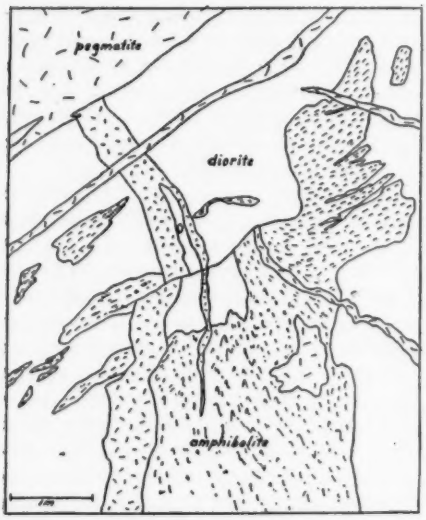

FIGURE 19.—DIORITE AND PEGMATITE CUT THE AMPHIBOLITE
½ mile south of the mouth of Grizzly Creek.

sills, which are younger than the effusives but older than the intrusives, are common in the area.

Contacts Between the Intrusives and Their Wall Rocks

The contacts between the intrusives and their wall rocks are well exposed at many localities. There are three types: (1) contacts with extensive brecciation suggesting forceful emplacement of granodiorite, (2) gradational contacts with migmatitic structures, (3) sharp contacts, either concordant or discordant.

Where the contacts are concordant they may belong to group 2 or 3, whereas discordant contacts are usually either of type 1 or 3.

The best example of contact breccia occurs near Grizzly Creek along the North Fork of Feather River. Angular blocks of the country rocks are cemented together with the granitic material (Pl. 5, fig. 2). The hornblende in the blocks has recrystallized, occurring in larger grains than in the amphibolite on the south side of the brecciated zone. To the north, the grain size of the hornblende and that of a part of the plagioclase grows. At this stage the blocks resemble coarse gabbro, and material between the blocks is still fairly light-colored. The second phase shows increasing of light minerals in the blocks and gradual disappearance of the large hornblende crystals. First the large hornblende crystals are scattered, occurring in swarms in places where formerly there were amphibolite blocks. These scattered large crystals become gradually fewer and smaller until the rock looks homogeneous. Diorite No. 118, was taken from this homogeneous contact phase. Thus every gradation from the dark medium-grained amphibolite to coarse gabbro and diorite can be seen among the angular fragments of the wall rocks; and further, the gradual disappearance of the outlines and darker color of these fragments when their "granitization" is completed.

According to the "reaction principle," the basic crystallization differentiates of the magma cannot be assimilated by the later, more acidic magma, since it is saturated in those minerals which were separated at the earlier stage. However, the gradual disappearance of the large hornblende crystals in the recrystallized blocks as well as the final homogenization of the contact rock lead to the conclusion that the Fe and Mg atoms must have been mobile. Abundant biotite appears in the homogeneous dioritic rock and is probably a reaction product between the hornblende and potash-bearing magma. Thus reaction between the older rocks and the magma as well as migration of the elements are responsible for the dioritic border facies.

Above and on the south side of the breccia in figure 2, of Plate 5, numerous white veinlets cut the fine-grained amphibolite at any angle indicating that the country rock was shattered also in this outer zone (Pl. 5, fig. 1).

Dioritic and pegmatitic dikes and small intrusions cut the amphibolite about half a mile south of the mouth of Grizzly Creek (Fig. 19). Limestone beds ranging from 20 to 80 cm. thick occur in amphibolite just east of this contact. Abundant epidote has crystallized at

the contact between limestone and amphibolite. Pegmatite veins cut across both. Larger angular amphibolite inclusions occur in granodiorite about 1 mile north from Grizzly Creek, and dark schlieren about half a mile further north.

An intrusive contact between gabbro, diorite, and granodiorite occurs north of Storrie. A large diorite intrusion occurs between Tobin and Storrie, bordered by gabbro south of Tobin and by amphibolite north of Storrie. Amphibolite is a remnant of the wall rock and contains some diopside in addition to hornblende and plagioclase. Gabbro on the south side of this remnant contains inclusions of this wall. Granodiorite to the south of this rather small gabbro forms a contact breccia with the gabbro, indicating clearly that the granodiorite is younger and that the gabbro was solidified before the intrusion of granodiorite occurred (Fig. 11).

The concordant contacts of the Merrimac pluton are sharp. The granodiorite shows a foliation parallel to the contact. The sharp, smooth, concordant contacts are probably due to the easy way that the intrusive penetrates the wall rocks parallel to the old structure. In the discordant parts of the contact, however, brecciation took place and dikes from the intrusive extend into the wall fairly far from the actual contact. This type of contact occurs 2 miles west from Coyote Gap. Thus the brecciation can be seen only where the resistance of the wall was greater. The broken fragments of the walls along the concordant contacts were either brought up into upper level, sank down into the magma, or were digested by it. The two latter possibilities appear more probable. The fragments that broke off along the concordant contacts probably had shapes of long lenticular bodies such as are still seen along the contact, 2 miles northeast of Brush Creek. At most localities the country rocks consist of amphibolites, serpentines, or other rocks with higher specific gravity than that of granodiorite. Similar but still attached parts of the wall rocks occur in many localities. On the other hand, the rather thin sheets torn out of the walls were more easily digested than the square blocks along the discordant contacts.

Clear signs of this kind of digestion of wall rocks in large amounts can be seen along the western contact of the Bald Rock pluton. The outcrops along the old Oroville-Quincy road just west of Lake Madrone consist of light bluish-gray granite (Sp. 408). The country rocks consist of metarhyolite, metadacites, metabasalts, and metamorphosed tuffaceous sediments. Long lenticular, partly granitized inclusions of these rocks occur along the above-mentioned road 1½ miles north of Lake Madrone. The contact west of the lake is slightly discordant so that the granite changes over to a metarhyolite and metadacite toward the south along a zone parallel to the structure. The granite shows a well developed foliation parallel to the same structure in the country rock. Thin section study shows that granite consists of plagioclase (An_{18}), quartz, some microcline, biotite, magnetite, and epidote. Quartz grains, showing strain shadows, are elongated parallel to the structure. The main part of the plagioclase occurs in rounded grains of medium size containing dust-like inclusions in their centers. Small plagioclase crystals, together with small quartz and orthoclase grains, fill the interspaces. In addition, there are large rounded or square plagioclase grains which are full of small sericite and epidote inclusions. These plagioclase phenocrysts are in many places aligned parallel to the foliation (Pl. 3, fig. 4) and are similar to those occurring in metavolcanics and tuffaceous sediments of the same area. They are thus most likely relicts, and this contact phase of the pluton can be considered to be a granitized dacite tuff.

South of this locality, a contact migmatite is exposed along the Bald Rock road (Loc. 206). The country rock here is a porphyritic metadacite. It occurs as an older portion in the migmatic rock just outside of the contact. The veins are granitic. In the field, this rock looks like an injection gneiss in which two thirds of the material consists of granite veins and a third of light-gray schist. Under the microscope, streaks of green hornblende and rows of altered plagioclase phenocrysts provide clear proof of the volcanic origin of the medium-grained, light-gray bands.

Contact migmatite occurs also north of the Headlands Mill on the Milsap road (Loc. 264). The coarse granite is here in contact with hornblende schists and metabasalts.

The occurrence of a large diorite west of Milsap Bar may be due to granitization of basic volcanics. There are clear signs of granitization along the contact south of this diorite along the South Branch of the Feather River (Loc. 250). The country rock consists of a dark amphibolite. Inclusions of this wall rock are common in the granodiorite near the contact. The granodiorite itself is a coarse-grained, brownish-gray rock in which plagioclase occurs as large individual grains or groups of grains. Thin-section study shows that there are abundant small epidote prisms included in these hypidiomorphic plagioclase crystals. Groups of strained quartz grains of medium size together with hornblende, biotite, and epidote fill the interspaces. Some sphene and a few orthite grains occur as accessories. To the west this rock grades gradually over to granite typical of the Bald Rock area (analysis 204) and to the east to amphibolite.

In the country rock to the east, the light minerals occur in separate large grains or groups of grains and their number decreases away from the contact. About 10 meters east of the contact, a granitic vein in the amphibolite shows gradational contact similar to the main intrusive and gives the impression that the vein itself was formed by granitization. Plagioclase and quartz in the wall rock crystallized as scattered grains or accumulations of grains from the material coming out from the granite intrusion. This phenomenon is very similar to that described from southwestern Finland where amphibolites were locally granitized (Hietanen, 1946, p. 1072).

When the individual plagioclase grains start to crystallize in the amphibolite, biotite appears together with the recrystallized hornblende. Thus, both potash and soda were introduced, but the potash reacted with the hornblende, forming biotite, and only plagioclase crystallized as large euhedral crystals. The advancing granitization changes the amphibolite first to diorite in which the main minerals are plagioclase, quartz, biotite, and hornblende, then to granodiorite, and finally to granite poor in potash feldspar or even to trondhjemite. Where the older rocks consisted of metarhyolites such as certain layers south of Lake Madrone, the "intrusive" contains considerable potash feldspar. In some other localities abundant epidote occurs as small prisms included in large blocky plagioclase crystals and as independent grains among the biotite, hornblende and quartz (Loc. 253). In this case, the country rocks were rich in epidote. Thus the influence of the digested older rocks can clearly be seen in the mineral composition of many contact facies of the plutons, while the centers are more homogeneous.

The inclusions of the wall rock in the granodiorite are granitized in a similar way. An addition of feldspar in larger grains gives the rock an appearance of a porphyritic diorite.

Mechanism of Intrusion

The internal structure and contact relations of the Merrimac granodiorite pluton illustrate the mode of emplacement of the granitic batholiths of the area studied.

The border zones of this pluton consist of granodiorite similar to that of analysis 18. The amount of microcline and quartz increases toward the center, which consists of a coarse-grained structureless granite-trondhjemite (No. 312) in which biotite occurs in large flakes and hornblende in short columns. Foliation is well developed in the contact zones especially near French Creek and at the mouth of Mountain Creek, but even in these localities it disappears within a short distance (200–500 meters).

The granite-trondhjemite seems to occur as a separate small dome in the center of the granodiorite intrusion, which dips steeply also in the central part of the pluton. An outcrop 2½ miles west of Merrimac along the road leading to French Creek shows an intrusive relation between medium-grained diorite and coarse-grained granite-trondhjemite (Fig. 20). Outcrops on both sides consist of granodiorite. It seems that even the rather homogeneous Merrimac pluton was not emplaced in one phase but rather that there were at least two: one granodioritic and another granitic-trondhjemitic. Granite-trondhjemite is clearly a slightly more acidic phase of the same magma as the granodiorite, and was emplaced shortly after granodiorite. The structure of the older rocks indicates that the pluton was emplaced in a tectonically suitable place: in the crest

of a large fold. How was the space obtained? This question cannot be answered directly, but the structure and contact phenomena give some suggestions. Movement in the magma must have been very slight, otherwise the inclusions would be deformed. Common brecciation of the walls indicates that stoping was one of the possible means by which space was obtained. The obvious swelling of the Merrimac pluton toward the north and south as well as general doming and buckling of the folds in the country rocks on the west and southeast side of the pluton suggest that magma pushed the country rocks aside. Finally the clear signs of digestion and granitization of the walls and fragments torn out of them show that these processes were not unimportant in furnishing more space. None of these three phenomena alone could have furnished enough space. This becomes clear if we assume that the Merrimac pluton took its place forcefully by pushing the surrounding rocks aside. This kind of mechanism of intrusion would cause an additional deformation of the adjacent older rocks. The compression would be perpendicular to the contacts and would result either in an extensive brecciation or, at a lower level and a higher temperature, in platy structure which would parallel the contact and would occur in the intrusive as well as in the country rock. This is true only along concordant contacts. In those localities where the granite cuts the older rocks discordantly, the granite is either structureless, shows a weak linear or platy structure parallel to the similar structure in the country rock, or there is a steeply dipping platy structure parallel to the general contact. The country rock, moreover, shows little signs of a second deformation parallel to the contact. The general parallelism of the contacts and the structures of the older rocks are therefore more likely due to the fact that the intrusion followed the easiest way, the pattern of the older rocks, when it was emplaced. Larsen (1948) has discussed the possibility of the plutonic rocks of Southern California forcing their roofs upward and finds good arguments against it in general lack of horst structures in the area. The same argument can be brought up in the present area. Ernst Cloos (1936) has suggested as one possibility that the magmas filled the open spaces in a similar row of "federkluftartiger Narben" as found in the fault zone on the east side of the Sierra Nevada. In the present area, the rounded form of plutons speaks strongly against the intrusion along a

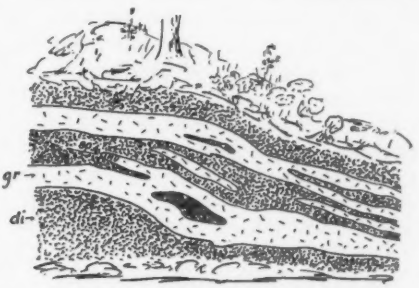

FIGURE 20.—GRANITE VEINS IN DIORITE
Inclusions in the vein are diorite. 2 miles west southwest of Merrimac.

fissure zone, and it becomes clear from the map that a portion of the older rocks is missing.

The general fragmentation of the walls and lack of strong flow structures favor the idea of stoping as suggested by Larsen (1948) for the batholiths in Southern California; but where are the stoped blocks of the roof? Are they far below the present exposed surface or are they digested by the magma? It is hard to believe that all the remnants of the roof had sunk while the rounded inclusions of approximately the same specific gravity were obviously floating. This would be understandable only if we assume that the older sedimentary and volcanic rocks were completely assimilated and only the fragments of the younger dike rocks, gabbros, and diorites corresponding to the basic forerunners remained as inclusions in the hotter center of the intrusive.

Digestion of the walls and brecciated blocks by the magma is clearly seen locally along the walls. But if we assume that magma rose passively and took its place by assimilating the roof (the word "assimilation" is used here to indicate similar phenomena to those which caused the homogenization of the contact breccia) then we would not expect magmatic structures but only relict structures in the igneous rock. In addition one would expect that the chemical composition of the igneous rock would

vary according to the digested older rocks more than it does. General platy structure parallel to the contact indicates that magma must have been moving while crystallizing, or that crystallization took place under a pressure directed toward the roof and the walls.

The force of the slowly moving magma was great enough to cause brecciation and to displace the broken blocks. Moreover, it was great enough to cause doming of the roof and curving of the walls parallel to the old structure. Pressure of the magma was directed not only toward the roof but also toward the walls as one would expect in a more or less liquid magma upon which a force is imposed. But this force was not great enough to produce all the space needed by pushing the country rocks aside. Stoping and assimilation took place durng the slow upward movement of the magma. General shattering of the roof and the walls by moving magma favored stoping. Finally, the physico-chemical conditions were suitable for assimilation of the stoped blocks. Granitization of the walls took place along the contacts of the granitic batholith. Several bodies of the magma rose serially. The contact relations between the various magma types show that the intrusions began with the basic members and became acidic towards the end. Since gabbro fragments are included in diorite and granodiorite, the gabbros were consolidated before the time of the intrusion of these more acidic magmas.

Metamorphism

Introduction

The sedimentary and volcanic rocks were metamorphosed regionally contemporaneously with the folding. Later a second (contact) metamorphism took place in the surroundings of the intrusives. This new metamorphism was most intense next to the contacts of the granodiorite intrusions but decreased just away from the contacts.

Regional Metamorphism

The rocks outside of the contact aureole of the intrusives show a metamorphism in epidote-amphibolite and green schist facies. Green amphibole and epidote together with either albite (An_{5-10}) or oligoclase (An_{10-30}) are the common association in the metadacites and metabasalts. Chlorite, occurring locally in minor amounts in these rocks, is in many cases due to later alteration rather than an indication of a low-grade metamorphism. In contrast to these, the slaty schist in the southwestern corner of the mapped area clearly belongs to the green schist facies. The rocks of similar composition near Pulga, near the mouth of the Little North Fork and elsewhere, consist of muscovite-biotite or hornblende schists belonging to the epidote-amphibolite facies like the metavolcanics with which they occur. At many localities, the fine-grained amphibole-epidote-chlorite schists are interbedded with amphibolites and mica schists indicating that the equilibrium was not attained during the metamorphism.

Primary textures in the volcanic rocks are well preserved. The rhyolitic flows especially seem to have been resistant to alteration and deforming movements. In many localities, only a few of the euhedral quartz and plagioclase phenocrysts show signs of deformation. Soda rhyolite No. 147 appears megascopically as a young fresh rhyolite. Under the microscope, it appears rather unaltered as if only the glassy ground-mass were devitrified. Many of the dacites and basalts appear unaltered in the hand specimen but microscopic examination shows that they are completely recrystallized. This can be explained as follows: The primary minerals in the rhyolite were in equilibrium in that temperature pressure field under which the recrystallization took place, that is, in the epidote-amphibolite and amphibolite facies. The minerals in the dacites and basalts were not in equilibrium and recrystallization took place. The anorthite component of the plagioclase altered to epidote. Light bluish-green, needle-like hornblende crystallized instead of augite and basaltic hornblende. Vesicles were filled by quartz and epidote. Where metamorphism took place under conditions corresponding to the green schist facies, the amygdules are filled by calcite and abundant chlorite occurs in the ground mass.

Contact Metamorphism

Crystallization of contact minerals like garnet, epidote, diopside, and vesuvianite are common in several localities along the southern contact of the Bucks pluton in the north, along the western, northern, and eastern contact of the Merrimac pluton and the northern contact of the Bald Rock pluton.

Best exposed and most easily accessible of these localities is the road cut near Grizzly Creek on the east side of the North Fork of the Feather River. The older rocks here consist of mica schists, amphibolites, and limestones.

Mica schists are transformed into migmatites with narrow white veinlets. Amphibolites are recrystallized as coarse-grained gabbros under the influence of added heat and granitic solutions. The limestones are represented by white and gray marble lenticles. The contact minerals occur along the contacts between the amphibolites and limestone lenticles. The following contact minerals occur in large quantities: garnet, epidote, diopside, hornblende, and tremolite.

Garnet is reddish brown and occurs in dodecahedrons ranging from 1 to 3 cm. in diameter. The index of refraction is $n = 1.81 \pm 0.02$. Green epidote with refractive indices $\alpha = 1.736 \pm 0.001$, $\beta = 1.759 \pm 0.001$, $\gamma = 1.772 \pm 0.001$, fills the interspaces. Many of the garnets show zoning. Alternating layering of garnet with epidote, quartz and iron-bearing dolomite is common. When the epidote-quartz-dolomite shell is removed, the garnet shows perfect faces of dodecahedrons.

Epidote occurs also, together with fine- to medium-grained, light-gray diopside, as euhedral dark- to pistachio-green crystals ranging from 1 to 10 cm. long. Some of the epidote crystals are almost black. Tremolite was found in long white fibres in the vugs between the diopside grains. It is clear that the occurrence of limestone and higher temperature near the contact has been responsible for the crystallization of these minerals.

The contact phenomenon in the amphibolite is seen in the larger grain size of the hornblende and in the granitic material between these large hornblende crystals. The result is therefore a rather nonhomogenous porphyritic gabbro which grades into diorite and granodiorite towards the intrusive.

The same mineral development, but on a smaller scale, occurs in the gorge of the Little North Fork, 2 miles southeast of Merrimac (Loc. 253). The limestone layers in amphibolite are transformed into layers consisting of garnet, epidote, diopside, and plagioclase. In the outer layer, the recrystallized hornblende is accumulated into small spots ranging from 5 to 7 mm. in diameter. Epidote fills the interspaces. Medium-grained diorite dikes cut this amphibolite at any angle.

Abundant light-green epidote occurs in streaks and lenticles in the gneissose trondhjemite along the eastern contact of the Bald Rock pluton (Loc. 260). The zone of granitization is here wider than in the other localities in the area and the epidote lenticles probably represent remnants of a limy layer in the older rocks.

Fine-grained apple-green, and translucent vesuvianite is included in serpentine and talc schist in a road cut about three-fourths of a mile north of Pulga Loc. 56,[1] thus half a mile south of the granite contact.

The crystallization of large andalusite holoblasts in the staurolite-bearing mica schists 1½ miles southeast of Big Bar Lookout is considered to be due to the contact metamorphism, as well as the crystallization of garnets in the mica schists 2 miles north of Merrimac (Loc. 129) and garnet and epidote in the biotite schist near the junction of the road going to Bear Ranch Hill (Loc. 127) and north and east of the Clear Creek Gold Mine (Loc. 126). This latter locality is about 1 mile south of the mentioned junction and near the contact of the granodiorite. Dioritic dikes cut across the wall rocks here. In the same locality, contact breccia is exposed along the old railroad leading from Coyote Gap to Merrimac.

These mineral assemblages indicate that the second metamorphism took place in the temperature-pressure conditions of higher amphibolite or pyroxene-hornfels facies. Similar relation between the earlier regional and later

[1] The locality was reported by Sterrett in 1910. The vesuvianite has been quarried and used as gem stone under the names californite and California jade.

contact metamorphism has been described by Durrell (1940) from southern Sierra Nevada.

Age of the Metamorphic Series

No fossil remains occur in the highly metamorphic schists and limestones of sedimentary origin. The lithologic correlation with other areas in Sierra Nevada leads to the conclusion that the age of the present rocks range from Carboniferous to upper Triassic or early Jurassic. The Triassic and Jurassic have been considered to be periods of volcanic activity (Lindgren, 1933). Because of this, many of the later writers believe that the thick volcanic formations were during this epoch. In the Inyo Range, fossil evidence was found that the limestones are of lower and middle Triassic age (Knopf and Kirk, 1918). These beds are overlain by thick beds of stratified volcanic and sedimentary rocks in which no fossils were found and which are considered by Knopf and Kirk to be upper Triassic in age. Hence, Erwin (1934) believes that the metamorphic rocks in northeastern Madera County belong to middle or lower Mesozoic, and Mayo (1934) considers the metavolcanics of southwestern Mono county to be Triassic. According to Macdonald (1941), the metamorphic formation between the Kings and San Joaquin Rivers is made up of a continuous series of interbedded sedimentary and volcanic rocks, all older than Mariposa. Definite age could not be determined. The main part of these rocks is believed to be Carboniferous with some older and younger beds. Metamorphic rocks south of the present area are grouped by O. E. Bowen and R. A. Crippe (1948) to sedimentary Calaveras formation (Carboniferous) and metavolcanics of uncertain age (Jurassic to Paleozoic).

The fine-grained isoclinally folded slaty schists near Las Plumas underlie and are older than the volcanics. Turner considered that they as well as rocks near Pulga belong to the Calaveras formation and are thus Carboniferous. No discordance was found between these rocks of sedimentary origin and the volcanics. The sedimentary series near the West Branch Bridge along the Feather River Highway contains conglomeratic beds, in which there are pebbles of fine-grained sericitic quartzite, mica schist, diabase, andesite and basalt. The conglomerates north of Pulga railroad station appear to be tectonic conglomerates in the tremolite-quartzite beds. The sedimentary beds near Pulga are overlain by a considerable thickness of stratified tuffaceous sediments, now represented by hornblende schists. A similar formation is exposed between the Merrimac and Bald Rock plutons striking from east to west. These two belts are lithologically identical and may belong to the same formation. The zone passing through Brush Creek can be traced 7 miles long and half a mile wide. It is cut off on the eastern end by granite intrusion; on the west, metabasalt and metarhyolite take its place In the northern zone, this stratum can be traced eastward from the North Fork of the Feather River at least 2 miles. The other formations have about the same magnitude, as indicated on the map. The metavolcanics occur as lenticular bodies of various sizes. Quartzitic beds together with mica schists are intercalated with the volcanic rocks and must be of the same age. Many localities in the area offer clear evidence of this; rhyolite No. 406 overlies the thinly bedded quartzite-mica schist formation, while near the mouth of the Little North Fork rhyolite No. 221 is overlain by sedimentary schists and quartzites. Intercalation of the sedimentary and volcanic rocks is best visible in a southeast-northwesterly zone, east of Merrimac. Small lenticles of limestones and thin layers of quartzites and mica schists are here interbedded with agglomeratic and tuffaceous beds. The metarhyolites occur as compact, less schistose lenticles ranging from 2 to 30 feet thick. The basaltic flows are more extensive, alternating with the agglomeratic and tuffaceous layers, now fine-grained highly schistose hornblende schists. Thinly bedded quartzite and bluish-gray mica schist together with small limestone lenticles near the junction of Deer Park and Gravel Range roads are indications of sedimentation during the period of volcanism. The same kind of alternation of sedimentary, tuffaceous, and extrusive layers can be traced eastward to the Sky High Road. The quartzite beds here are somewhat thicker but otherwise clearly of the same type.

The section along the railroad near Pulga (Fig. 5) gives a typical example of the sedimen-

tary formation in the main part of the area mapped by the writer. The area near West Branch Bridge differs from this greatly, the sedimentary formation being more homogeneous and consisting mainly of slates and conglomerates.

Serpentines were probably intruded into the bedrock complex contemporaneously with the folding. The contacts are as a rule concordant and the shape of the serpentine bodies is that of long lenticles. The border zones, usually represented by talc schists show a pattern of deformation similar to the wall rocks. In many localities, these serpentine bodies are cut by later ultrabasic dikes now consisting of the same kind of serpentine as the main body. Two examples of these younger serpentine dikes are described above, one from the railroad cut about three-fourths of a mile north of Pulga railroad Station and another from the road junction to Yankee Hill.

Age of the Granitic Intrusions

As in the main Sierra Nevada area, the granitic batholiths in the present area were emplaced after the folding and metamorphism of the sedimentary and volcanic rocks. The time of intrusion of the main Sierra Nevada batholith is known to be post-Mariposa and is considered to be late Jurassic or early Cretaceous by several authors (Knopf, A., 1918; 1929; Hulin, 1933; Mayo, 1934; 1935; Erwin, 1934). Hinds (1934) showed that the granites of the Klamath Mountains north of the area under consideration are probably very late Jurassic. According to Böse and Wittich (1913), Wittich (1915), Woodford and Harriss (1938), and Larsen (1948), the granitic rocks in Lower California are Upper Cretaceous.

Metasomatic Alterations

Best examples of metasomatic rocks in this area are the coarse gabbro and diorite about 2 miles south of Pulga. They are well exposed along Highway 24 and again along the railroad on the east side of the North Fork of the Feather River. The country rocks are amphibolites and serpentines and the contacts towards them gradational.

A photograph of a polished hand specimen of the metasomatic diorite is shown in figure 3 of Plate 5. The white mineral is mostly square plagioclase crystals. The dark mineral is a colorless to light-greenish hornblende. The contacts with the country rocks are irregular and abundant narrow veinlets of white material occur in them. This white mineral appears to be almost pure zoisite. Its chemical composition (Table 4) is calculated from the rock analyses 12 (Table 1). Zoisite occurs in long prisms which in thin section show gray-bluish interference colors. The optical properties are $\alpha = 1.702 \pm 0.002$, $\beta = 1.703 \pm 0.002$, $\gamma = 1.708 \pm 0.002$; $+ 2V = 10°$. Optic plane $\perp$ (010) cleavage. Some plagioclase and a few hornblende needles included in plagioclase occur as minor constituents in these white veinlets. The chemical and mineralogical compositions of the veinlets is shown in table 1 (No. 12). Chemical analysis of the gabbro situated between the diorite and amphibolite is shown under No. 11 in the same table. No. 10 represents a transition zone between the gabbro 11 and amphibolite with white veinlets. Minerals in the gabbro (Pl. 5, fig. 4) are plagioclase (An_{78}) with irregular twinning lamellae, light-green to colorless hornblende, some epidote, and chlorite. The same minerals appear in the amphibolite but there is more plagioclase and less hornblende in the gabbro. In the central part of the occurrence the amount of plagioclase is considerably greater (Pl. 5, figs. 3, 4). Thus it seems to have been pure plagioclase (An_{78}) which was added into the amphibolite

Table 4.—Chemical Composition of a Zoisite From White Veinlets Two Miles South of Pulga

Calculated from analysis 12, Table 1.

	Weight %	Mol.
SiO_2	38.63	0.6406
Al_2O_3	34.44	0.3370
Fe_2O_3	0.77	0.0048
FeO	0.09	0.0013
CaO	22.95	0.4092
H_2O	3.12	0.1732
	100.00	

by metasomatic processes. Zoisite instead of anorthite crystallized in the veinlets from the same material because more water was available and probably also because of lower temperature.

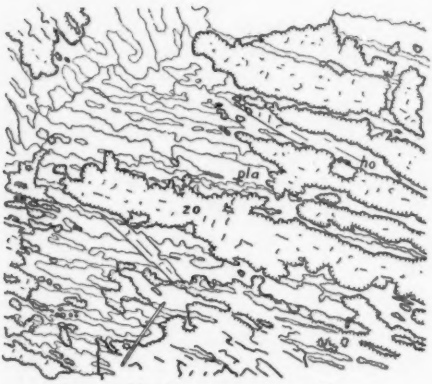

FIGURE 21.—REPLACEMENT OF ZOISITE BY PLAGIOCLASE

In the border zone of the metasomatic gabbro pla = plagioclase, zo = zoisite, ho = hornblende. Note the wavy boundary lines between the twinning lamellae in the plagioclase. 2 miles S. of Pulga, Loc. 28. 89×

The introduction of plagioclase took place after the folding of the older rocks since the affected areas occur as irregular lenticles in the serpentines and amphibolites. The contacts cut across the structures of these older rocks. Individual plagioclase grains, groups of grains, and narrow veinlets occur outside of the metasomatic gabbros and diorites. The two smaller diorites along the railroad, 2 miles south of Pulga, occur in serpentine in a similar way and are extensions of the larger metasomatic gabbros and diorite occurrence described above.

Long light-green needles of hornblende and fine-grained zoisite replace antigorite along the outer zones of this occurrence. Then plagioclase starts to appear first as separate grains and groups of grains. The serpentine grades thus gradually into hornblende gabbros and, with increasing amount of plagioclase, into diorite. The total amount of magnesium is greatly reduced and the amounts of aluminum, calcium, and alkalines increased during the metasomatism. The amounts of silica and iron are about the same. Thus when the plagioclase-bearing solutions invaded the serpentine, hornblende crystallized out as a reaction product between serpentine and the solutions rich in calcium and aluminum. Zoisite instead of anorthite crystallized out at lower temperatures in the presence of abundant water freed in this reaction. Magnesia was removed in the solution. Plagioclase (An_{78}) crystallized in the center of the invaded area. The temperature here must have been considerably higher to allow the crystallization of laboradorite instead of zoisite and albite. The center of the invasion was most likely the area where the dioritic rock occurs and the transition zones towards country rocks represent zones with smaller amounts of material added and also a smaller amount of heat brought into the rock. The phenomenon is comparable to the granitization but, instead of alkali feldspars and quartz commonly added into the rock during the granitization, here a rather basic plagioclase was deposited from the solutions.

The source of these solutions may be in some basic igneous rock, or they may have been derived from the older metavolcanics by differential resolution. Epidote in the metavolcanic rocks seems to have been very mobile. Narrow veinlets and small accumulations of green epidote are extremely common in the metavolcanics. Amygdules usually consist of quartz and epidote. Occurrences of zoisite instead of clinozoisite or common epidote may be due to the crystallization of abundant hornblende as a reaction product between serpentine and epidote-rich solutions. Dehydration of zoisite gives rise to the crystallization of a calcic plagioclase. Indeed, this crystallization of plagioclase at the expense of fine-grained zoisite can be seen in thin section (Fig. 21). First, irregular lamellae of plagioclase appear in massive, fine-grained zoisite. The amount of zoisite decreases, giving rise to the other set of twinning lamellae. The boundry lines between these two sets appear to be fringed resulting in a peculiar twinning (Fig. 21) typical of these metasomatic rocks.

GOLD-BEARING QUARTZ VEINS, HYDROTHERMAL ALTERATIONS, AND ORE DEPOSITS

Most of the gold-bearing quartz veins occur in the country rocks near the granite and

granodiorite batholiths. Many of them occur in the actual contact zone or even in the granite. The width of the veins ranges from a few cm. to 1 meter, rarely more. The gold occurs mainly in a clay-like material which fills a narrow shearing zone in the vein or on one side of the vein. Vein quartz shows striation against the clay-like material. Gold may be found also as minute flakes together with the pyrite between the quartz grains especially in the drusy vugs which are numerous in the veins and are lined by quartz crystals. This mode of occurrence is largely the same as described by Knaebel (1931) in the Mother Lode district. According to Knaebel, gold was precipitated in open spaces and came along planes of movement in the pre-existing quartz veins in the Grass Valley district. He gives the following sequence: (1) introduction of quartz and pyrite; (2) crushing; (3) introduction of gold, galena, and sphalerite. According to the miners, the gold ore occurs in 'pockets' in these narrow veins and large distances between these 'pockets' consist of plain gangue material with just a trace of gold. Many mines, once productive, are still operating on a small scale because the miners always hope to hit another 'pocket'. The only larger-scale gold mining is in the Big Bend area, the Surcease Mine of Butte County. This mine is operated by Hoefling Bros. and its production in 1945 was 540 fine ounces (Bur. Mines, Mineral year book, 1945).

Gold is mined from a quartz vein which occurs between porphyritic trondhjemite (analysis 7, Loc. 328) and fine-grained chlorite-tremolite schist of tuffaceous origin. Both the trondhjemite and the schist are hydrothermally altered. Addition of potash in the form of abundant muscovite is apparent as is an abundant addition of pyrite. Calcite is the third hydrothermally introduced mineral in the country rocks. Chloritization of biotite and alteration of the anorthite component of the plagioclase to epidote in the trondhjemite are further signs of the hydrothermal activity.

A. Knopf (1929) has described hydrothermal alteration of the country rocks of the gold-bearing quartz veins in the Mother Lode district and pointed out the abundant addition of carbonate in the veins and country rock.

Ingerson (1940) concludes that the ore-bearing solutions must have been dilute and not a 'ore magma' consisting of gangue material plus water, since carbon dioxide decreases the solubility of silica in pure water and in solutions containing alkalies.

Copper and zinc were taken from the Big Bend mine during World War II. Location of this mine is about 2 miles east of the Surcease Gold mine. The country rocks are a fine-grained dark soda-rhyolite and a dacite tuff both metamorphosed in the epidote-amphibolite facies and later hydrothermally altered near the ore body. Abundant pyrite cubes occur in the tuffaceous material, which consists of quartz, plagioclase (An_{20}), chlorite, and epidote. The small plagioclase laths are randomly oriented. Schistosity is well developed in this tuffaceous layer but lacking in the soda-rhyolite.

The ore minerals are chalcopyrite, sphalerite, and galena. Some specimens contained up to 35 per cent copper, but the ore body was rather small. Production of the mine operated by the Hoefling Bros. in 1945 was: 2,502,000 lb. zinc; 48,000 lb. lead; 390,000 lb. copper, and 18,750 fine ounces silver.

Chrome ore was quarried on the east slope of the North Fork of Feather River just north of the highway bridge near Pulga. The mineral chromite occurs as syngenetic segregations in the serpentine.

Mineral Development in the Effusive and Intrusive Rocks and Their Genesis

In earlier papers, the albite in the keratophyres was thought to be a primary constituent due to a high content of sodium and low content of potassium in the magma from which keratophyres were crystallized. Gilluly (1935) gives a great number of references to keratophyres and discusses the problem in connection with spilites. He suggests that both rocks may have a common origin. He also points out that most of the known keratophyres are associated with marine sediments as are also most of the spilites. Nolan (1930, 1935) states that albite in the altered andesites and trachytes of the Tonopah mining district is a result of later introduction of soda. Thus the sodic plagioclase phenocrysts would be a result of the replacement of either more calcic plagioclase or

orthoclase depending on the original rock being either basalt or andesite, or on the other hand either rhyolite or trachyte. The mode of replacement could be either autolysis (Eskola, 1925) or regional albitization.

The author cannot see any reason to assume that the plagioclase phenocrysts in the present area have been a result of albitization. The mineral development in the biotite-rich metarhyolite No. 406 (analysis 5) shows a ratio K_2O:Na_2O to be the same as in normal rhyolites but there is no potash feldspar present in the rock. The texture of this rock shows no signs of deformation or alteration except recrystallization (or devitrification) of the groundmass. The perfectly hexagonal biotite plates must have crystallized out from a liquid magma. Absence of potash feldspar in this rock is due to the abundant crystallization of biotite.

Most likely the differentiation followed a pattern similar to that suggested by Goldschmidt (1916) for the trondhjemites. Abundant biotite crystallized at an early stage from a water-rich magma. This used potash to such an extent that no potash feldspar can crystallize in the end member. In the present area, the state when abundant biotite was crystallized is represented by analysis 5 (Table 1). This rhyolite (No. 406) differs mineralogically from the other extrusives of this region mainly because of its high content of biotite. It does not represent an independent magma but an accumulation of biotite in a magma of rhyolitic or rhyo-dacitic composition. The albite and quartz phenocrysts are few and small. Crystallization of these minerals was either delayed because of the high content of water in the magma, or they floated and were separated from the magma which carried abundant biotite. The specific gravity of the magma was lowered and its viscosity greatly decreased by the high content of water. The structures indicating turbulent flow in the rock 406 support this view. The biotite phenocrysts probably were able to sink more rapidly in this kind of magma than in the normal rhyolite magma. The separation of the main part of biotite did not take place at the early stage as suggested by Goldschmidt for the Norwegian trondhjemites, i.e., at the dioritic stage, but in the rhyolite magma itself. Beskow (1929) has described soda-rich acidic effusives from Sweden and considers their peculiarity to be due to the crystallization differentiation in the same manner except that biotite in the Swedish effusives began to separate even earlier than in the Norwegian trondhjemites. According to the author's investigation in southwestern Finland (Hietanen, 1943), abundant biotite was separated at the dioritic as well as at the diorite-trondhjemitic stage.

Many metabasalts in the present area contain some chlorite which is clearly an alteration product after biotite, and as a rule metadacites contain either biotite or muscovite. Muscovite may have been introduced hydrothermally, but in many specimens it occurs in a manner which suggests that it crystallized at the expense of potash originally present in the groundmass. For instance the dacite porphyry, No. 140, near the mouth of Berry Creek contains both 'primary' and secondary muscovite.

Thus, the metamorphic soda-rhyolites of this region are most likely differentiates from the same parent magma as the metadacites and the biotite-rich metarhyolite. The amounts of orthoclase and plagioclase vary geographically; all flows on the southern part of the mapped area are rich in sodium, whereas the normal rhyolites are found in the northern part of the area. Both varieties together with the intermediate types occur in the eastern part. As a rule dacites are associated with the soda-rhyolites.

The occurrence of quartz porphyries and an intrusive equivalent of the soda-rhyolites (trondhjemite 328) further supports the viewpoint that the soda-rich effusives were crystallized from a soda-rich trondhjemitic magma. The rocks of this trondhjemitic suite are characterized by a high content of sodium, calcium, and iron and low content of potassium and magnesium.

Thus, we seem to have a splitting of the basaltic magma into two branches, as follows:

1) normal basalt → rhyolite suite
2) basalt → dacite → soda-rhyolite (trondhjemitic) suite

The chemical peculiarities of the latter series can be traced to the metabasalts (Analysis 2). The most striking difference between it and

normal basalts is in the amounts of magnesium and calcium. Figure 15 shows that the metabasalt No. 156 has the same chemical characteristics as the dacites and soda-rhyolites; the percentage of MgO is much lower and the percentages of CaO and FeO higher than one would expect.

The basic forerunners like uralite porphyry No. 175 contain abundant chlorite in addition to green hornblende as a dark mineral. The chlorite is probably an alteration product after biotite and many of the larger hornblende crystals show outlines of augite, suggesting that this rock was originally a biotite-bearing augite porphyry. Thus, the scarcity of potash feldspar in the intrusive series may have at least in part the same explanation as that in the extrusives and in the trondhjemitic suite in general. The mineral development in the intrusives shows that the magma contained a considerable amount of water. Hornblende and some chlorite or biotite are the dark minerals even in the most basic members. Even though diopside is a common mineral in the metamorphosed wall rocks near the contacts, indicating a high temperature in the contact aureole during the intrusion, it is rarely found in the intrusive side. The inclusions of ilmenite in the centers of many hornblende grains suggest that the hornblende crystallized during the latest phase of crystallization at the expense of titaniferous pyroxene. Epidote is a common additional constituent occurring in fairly large crystals together with hornblende and biotite. This occurrence suggests a primary origin and can be considered as a further indication of a high content of water in the magma. Another important factor was a reaction between the hornblende-bearing older rocks and the magma. The presence of the potash feldspar in the pegmatites indicates that the original magma contained considerable potash. Biotite crystallized out at the expense of this potash and the hornblende present in the stoped blocks of the roof and the walls. The potash reserve of the magma was used up in this reaction to such an extent that no potash feldspar could crystallize in the contact facies and that its amount in the center of the pluton was greatly reduced.

It is natural to draw a conclusion of a genetic relationship between these two series on the basis of the chemical and mineralogical identity. This identity could be considered to offer a proof for the metamorphic origin of the intrusive series. However, the contact relations as described above show that the intrusives were emplaced by injection from below rather than formed by metamorphic or metasomatic processes from the older rocks at the present level. The assimilation of large amounts of stoped blocks of the country rocks probably modified the intrusive magma, or both of these magmatic series may originate in the same source, the chambers being in a lower level than now exposed. Importance of the metasomatic processes can be traced along the contacts of the intrusions in several localities. This is especially true if the intrusive rock is acidic as in the Bald Rock pluton. The higher content of silica in this particular intrusion is not due to larger quantities of older rocks (sedimentary) granitized. Rather the high acidity of the magma was preserved in spite of great quantities of basic rocks (volcanic) digested. A wide border facies here was formed by granitization. The solutions which caused the granitization contained silica and the other components of feldspars and their source is to be sought in the same igneous magma from which the main part of the granite pluton crystallized.

However, the igneous granite now exposed may have been formed by remobilization of a metasomatic granite formed elsewhere, and we have to find the origin of the first granitizing solutions, emanations, and ions. Was it a granite magma, basaltic magma, or ultra metamorphic rock? As a rule, the basic intrusives do not have granitized or migmatized contact aureoles, which make the basaltic magma less probable as a source of granitizing material. The composition of the material added to the older rocks during their granitization is close to that of the granite pegmatites and it may originate in juveuile granite magma or in metamorphic rock through selective resolutions and metamorphic differentiation. This pushes the problem back to a lower level than is now exposed in the area under discussion. If we assume that the deeply eroded parts of the earth's crust, the Precambrian areas give a

true picture of the lower level in the younger mountain chains, we must admit that metamorphic differentiation, granitization, and palingenesis play an important role in making any granite. On the other hand, the observations in the area under consideration offer proof that acidic magmas are differentiation products of more basic magmas. Granitization and assimilation of the country rocks were found to be important in adding to the total amount of granite. Thus, the question does not seem to be whether a certain granite is igneous, metasomatic, or metamorphic, but rather what was the role of each of the possible granite-making processes during its formation: (1) crystallization differentiation, (2) assimilation, (3) granitization, (4) ultra metamorphism, (5) metamorphic differentiation, or (6) palingenesis. In the Precambrian areas, regional granitization starts with metamorphic differentiation and metasomatic addition of feldspars and quartz, and is followed by homogenization and locally by mobilization. In the present area the granitization is restricted to the contact zones and it is primarily a metasomatic replacement of hornblende by biotite, plagioclase, and quartz. Digestion of older rocks was important in modifying the magma and adding to its total amount. The calculated mean composition of the older rocks is close to that of gabbro. If we add one portion of these older rocks to two portions of granitic-trondhjemitic magma, chemically similar to the rock No. 204, we would have a rock of granodioritic composition, and the space furnished by the assimilation would be about one third of the total volume of granodiorites. Metasomatic processes which involve expelling or fixation of Mg and Ca are common in the older rocks, but such regularity as to suggest a basic front (Reynolds 1947) around the granitized aureoles is questionable. Rather, Mg and Fe were incorporated to the intrusive rock. However, more Mg may have been fixed in the serpentinization of peridotites and tremolite rocks than was freed from the olivines during their serpentinization. Only the addition of potash is necessary for the formation of biotite at the expense of hornblende near the contacts.

Table 5 shows a comparison of the possible granite-making processes in a Precambrian area in Southwestern Finland (Hietanen, 1943; 1947) and in the area under discussion.

The mechanism of metasomatic and metamorphic processes is not discussed in this paper. The writer believes that experimental work would give a better picture of this than field observations, whose interpretation depends on the prejudice of the observer. The recent work by Jagitsch (1949 a; 1949 b) suggests that only a minor role should be given for the diffusion in solids during metasomatic introduction of elements over great distances (Jagitsch, 1949 b, p. 101). Especially, if abundant quartz is added (as during the granitization of older basaltic effusives) liquid phase should be assumed, since the diffusion of silica in solid is much slower than that of alkalies.

Sequence of Events

The oldest rocks in the area are the sedimentary series near Las Plumas, near Pulga, and probably part of the sedimentary series in the Milsap Bar and Marble Creek areas. These schists are correlated with the Calaveras formation which is considered to be Carboniferous in age but may include older beds and also younger layers.

A thick series of tuffs, agglomerates, and lava interbedded with sediments overlie these older sediments. Because of a lack of fossils and because both formations are equally deformed and metamorphosed during the Nevadan orogeny, it is impossible to tell the difference between the sediments belonging to each of these two groups. Detailed tectonic mapping would be of some help, but the profound folding and deformation and the discontinuity of outcrops make this difficult. In addition, the area mapped for the present investigation is too small to give a stratigraphic picture. The ultrabasic rocks, mostly peridotitic and now represented by the serpentines, invaded the older formation contemporaneously with the folding. The hydration of these peridotites took place contemporaneously with the emplacement and partly shortly after it.

The intrusion of serpentines was followed by a series of dikes and sills emplaced when the compressional forces were still active but weakening. This series is represented by serpentine

TABLE 5.—COMPARISON OF GRANITE-PRODUCING PROCESSES

Precambrian Southwestern Finland (Hietanen, 1943, 1947)	*Jurassic* Merrimac Area
(1) Fractional Crystallization Differentiation (Bowen)	
Granitic rocks belonging to various intrusive series probably were in part formed by crystallization differentiation.	Evidence in favor of this hypothesis is found among the metamorphosed extrusive series. Intrusive series may have been in part formed in this manner.
(2) Assimilation. Incorporation of older rocks into intrusive granite is common in both areas. It was important in adding total amount of granite and to modify it, but assimilation alone could not produce granites.	
(3) Granitization. Mainly elements from which feldspars and quartz crystallize, are introduced into country rocks to the extent that they are transformed to granites. Remnants of older rocks offer proof of the origin of these metasomatic granites. Granitization, like assimilation, requires a source of granitizing solutions or elements.	
Granitization in regional scale is common. The granitizing solutions, emanations and ions may originate either in igneous granite magma or in metamorphic rock (see 5).	Granitization occurs along the contacts of granite batholiths. The source of granitizing solutions, vapors, and ions is the main igneous body.
(4) Ultra metamorphism may change only the rocks which have a granitic composition (sandstones, rhyolites) to granites. The sandstones themselves are weathering products of granitic rocks, so that the problem returns to the origin of the primary granitic rock.	
No direct evidence of granite being an ultrametamorphic rock was found, but this case has been described in literature from other highly metamorphosed areas.	No evidence.
(5) Metamorphic differentiation. At certain physico-chemical conditions, feldspars and quartz tend to separate from basic minerals and gather into veinlets, lenticles, and stocks, usually parallel to pre-existing s-planes. The mechanism is not known, but it seems to be greatly accelerated near granitic intrusions and generally part of the material in veinlets originates in the igneous granite.	
Metamorphic differentiation is an important granite-producing process.	No signs of metamorphic differentiation were found in the area under discussion, but this process may have been active at the lower level during the formation of granitic magma.
(6) Palingenesis. Granites formed by various processes described above and rocks of granitic composition may become mobile and act like true igneous granites. The palingenetic magmas are supposed to have the same properties as the true igneous magmas. Therefore, if the direct evidence of the metamorphic origin is missing, it is impossible to distinguish a juvenile magma from a palingenetic magma.	
Palingenetic granites are common, and do not seem to belong to any particular intrusive series.	No evidence was found, but as in (5), this process may have been important during the formation of granitic magma.
MIGMATITES	
Regional migmatization. Feldspars and quartz in the veinlets were derived from the host rock by metamorphic differentiation or introduced from a granite magma.	Contact migmatites. Granitic material in the veinlets is introduced from the main igneous granite magma.

dikes, gabbro intrusions and dikes, diabase sills and dikes, dacite porphyries, and trondhjemitic intrusives and dike rocks.

Emplacement of granodioritic batholiths followed these forerunners. In these intrusives, there are no signs of the type of deformation seen in the older rocks. The platy structure which occurs locally along the border zones is parallel to the contacts and is more likely either an igneous or a relict structure than a result of deformation after the emplacement. The igneous activity started with intrusions of small gabbroic bodies and continued by emplacement of large masses of granodiorite and granite-trondhjemite. The pegmatites, aplites, and basic dike rocks filled the cracks formed in the wall and the intrusive during and after the consolidation.

Quartz, calcite, and pyrite were deposited along the cracks formed during later faulting and finally the solutions carrying gold, galena, sphalerite, and pyrrhotite filled the open spaces, vugs, and shear planes in these veins.

The older rocks were metamorphosed in green-schist and epidote-amphibolite facies during the folding. The emplacement of granodioritic batholiths caused the second metamorphism in higher amphibolite and in pyroxene-hornfels facies along the contacts.

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Manuscript Received by the Secretary of the Society, May 1, 1950.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 609-618, 1 FIG. JUNE 1951

IMPROVED DITHIZONE FIELD TEST FOR HEAVY METALS IN WATER

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ABSTRACT

In many areas of the Pacific Northwest, heavy run off from rain, melting snow, and ice fields causes extreme dilution of the heavy metal content of natural waters.

Actual field work has shown that in many areas it is only practical to apply geochemical methods in the search for copper and zinc if it is feasible to detect as little as 0.001 p.p.m. of these metals in natural waters.

A modification of the dithizone methods which have already been described has been developed. This modified method is simple, as accurate as is necessary for field work, and can readily detect heavy metals in concentrations as low as 0.001 p.p.m.[1] Under favourable conditions, even lower concentrations can be detected by an experienced field man.

During the 1950 field season, we have started to refer to this modification in the use of dithizone as the "emulsion" technique, in contrast to the "classical" dithizone methods employed by earlier writers.

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INTRODUCTION

Examination of the metal contents of the natural waters in a neighborhood seems the most simple and obvious first step in any search for hidden ore by means of the geochemical dispersion of its weathering products. Vogt and Rosenqvist (1942) in Norway, Sergeev (1946) in the U. S. S. R., Huff (1948) in the U. S. A., Webb and Millman (1950) and Millman (1950) in Nigeria have all shown how heavy metal content of natural water can be used in prospecting. Our early efforts to apply these techniques in British Columbia met with failure. We can now explain this failure. There was not a sufficiently high heavy metal content in the water to obtain significant results with these methods.

In all the areas investigated in British Columbia, we have found that, except for streams nourished by artifical concentrations, such as in the vicinity of mine workings, the metal contents of natural water coming from buried mineralization is such that detection is beyond, or perilously close to, the limit of sensitivity of the methods described by earlier writers.

After many experiments we became satisfied that dithizone remained the key chemical for geochemical work. However, an improvement in technique was essential if it were ever to be used in the field by prospectors without special chemical training.

In this paper we describe a technique which, practice has shown, is simple enough for many prospectors and sensitive enough for most areas in British Columbia. In essence, the technique consists of using a aqueous solution of dithizone to react rapidly on the minute amounts of heavy metals in solution, and of collecting the resulting heavy metal dithizonates, together

[1] = parts per million.

with any unused dithizone, by means of an emulsion of some suitable organic solvent immiscible with water.

By one of those curious quirks which plague many researchers, it appears to be more difficult to duplicate marginal results in a laboratory than it is in the field. This may be caused by the inherent difficulties of maintaining an abundant supply of truly "pure" reagents; or a greater danger of oxidization in a laboratory; or salting by dust particles containing traces of heavy metals; or combinations of these and possibly other factors—we do not know. In attempting to duplicate a demonstration in our laboratories, we have always to consider the hazard of a stray gamma or two of metal being introduced unknowingly into our assay, a hazard less frequently encountered in areas remote from civilization. The high proportion of negative results obtained from "blank assays" shows that in general the reagents are pure enough

Acknowledgments

The authors express their deep gratitude to the Geological Society of America for grants from the Penrose Bequest which enabled them, with the help of the Department of Geology and Geography of the University of British Columbia, to complete the theoretical part of this research. They are especially indebted to Mr. W. G. Huber of the B.C. International Engineering Co. Ltd., Kennco Explorations (Canada) Limited, and Mr. E. C. Roper of Britannia Mining and Smelting Co. Ltd., for giving them the opportunity to test exhaustively this "emulsion" technique in the field during the 1949 and 1950 field seasons. Thanks are also due to Mr. and Mrs. F. Brent, Mr. Cliff Harrison, Mr. Bill Harrison, Mr. G. H. Lee, Mr. A. F. Killin, Mr. H. Currie and Dr. R. M. Thompson who graciously played the role of guinea pigs and associates in these first practical attempts to discover and trace ore by means of excessively diluted natural waters.

Chemical Considerations

Huff (1948) has given a good account of the use of the classical dithizone method for determining zinc and other metals in water; Lovering, Sokoloff, and Morriss (1948) and Reichen and Lakin (1949) have described practical uses for this reagent. We have already given a description of the gray end-point titration method, (Warren and Delavault, 1949).

The red complex formed by dithizone reacting on zinc may be diluted and be still visible until the amount of zinc present is only 10.10^{-4} milligram or 0.01 γ per ml. of carbon tetrachloride. Unfortunately, the efficiency of extracting a metal from water by shaking that water with a carbon tetrachloride solution of dithizone is far from being perfect. Huff used this method with an excess of dithizone, which he removed with dilute ammonia, and found the faintest color detectable corresponded to a content of 0.1 γ zinc or 0.5 γ of copper in 50 ml. of water.

Two of us (Delavault and Irish, 1949), with the help of a shaking machine which enabled us to make simultaneously five extractions under the same mechanical conditions, found that there was no special range of pH or concentration of reagent by which the extraction of copper could be effectively speeded up. Variations in the speed and duration of shaking do not bring about marked improvement of extraction, because dithizone itself may become oxidized to a yellow compound which does not react further with metals. We decided to consider the problem anew.

Metal ions are insoluble in carbon tetrachloride and, under acid conditions, dithizone is insoluble in water. Heavy metals in water and dithizone in carbon tetrachloride could react only by diffusion through the surface of droplets. Reaction should be made much easier by introducing dithizone in a finely divided state directly into the metal-bearing water. The addition of a solution of dithizone in dilute ammonia to acidified water answers the requirement. Metal-free water assumes a bluish color with an addition of free dithizone. Metal present in water can react with dithizone under these conditions and indeed, if added to water containing enough metal, dithizone turns a pale but distinct purple or red.

The addition of appropriate amounts of dithizone dissolved in a weakly ammoniacal solution to a solution containing more than 50

γ per liter of heavy metals results in sufficient coloration for a successful qualitative test to be made without further manipulation. However only after it has been dissolved in some organic liquid will dithizone assume the green color normally associated with this chemical.

Metal dithizonate can be collected with carbon tetrachloride much more speedily than the metal can be extracted by a solution of dithizone. However, in the case of water containing very small amounts of a heavy metal, this would mean adding to 100 ml. of this water as little as 12 γ of dithizone and collecting the resulting complex, plus any excess reagent, in one ml. or less of solvent. All this involves appreciable loss of organic solution because of smearing of the walls of the shaking vessel. This in turn necessitates careful cleaning with acid and this, added to the trouble of carrying such impedimenta as shaking funnels on traverses in rough country, made us anxious to find a way of avoiding shaking methods in the field. What seemed to us a simple and rational solution was to make in advance an emulsion of a solvent which would break up and disperse when added to water to be tested, having become unstable because of a change of pH, and would then come together again, having in the meantime collected both dithizonate and any excess dithizone. This entailed no special difficulty, most emulsifiers being alkaline, and zinc and copper being extractable from the acid solution. A solvent lighter than water was chosen in order that the droplets would rise, be more easily seen at the surface, and be more easily collected. Furthermore, a moderate volatility and a very low solubility in water were needed. Of all common chemicals, xylene seemed to best meet these requirements.

Reagents

(A) AMMONIACAL SOLUTION ($\frac{1}{50}$ AMMONIA): Take 2 ml. of pure concentrated ammonia and dilute with pure water, double distilled if necessary, to 1 liter. Keep well stoppered to avoid action of carbon dioxide from the air. Concentrated ammonia usually needs no special purification.

(B) DITHIZONE SOLUTION: Crystalline dithizone does not dissolve easily in dilute ammonia. Consequently, it is better to make up a stock solution of dithizone in carbon tetrachloride—60 milligrams of dithizone dissolved in 1 liter of carbon tetrachloride—and shake in a shaking cylinder equal volumes of this solution and of the $\frac{1}{50}$ ammonia which dissolves the dithizone. One must take care to make up only enough ammoniacal solution for the needs of the next few hours. The dithizone in the ammoniacal solution rarely keeps for more than a few hours, even under temperate climatic conditions. However, the carbon tetrachloride dithizone solution can be kept for weeks if placed in a moderately cool and dark place. One may take more ammoniacal solution than carbon tetrachloride solution if desired. We ourselves take equal amounts as this simplifies standardization and makes it possible to use either "classical" or "emulsion" techniques without worrying about any changes of value in the dithizone-bearing solution. A negligible amount of dithizone may remain with the carbon tetrachloride. However, in the field, we find it accurate enough to assume that, if 1.0 ml. of dithizone carbon tetrachloride solution neutralizes 3.8 γ of copper, 1 cc. of the ammoniacal dithizone solution will do likewise.

After the ammoniacal and carbon tetrachloride solutions are shaken, they rapidly separate and for all practical purposes remain immiscible. The upper layer is then drawn off as needed by means of a pipette. The lower layer may be yellow from oxidization products insoluble in ammonia or greenish because of a little unextracted dithizone.

Even under favorable climatic conditions, it is best to bring dithizone to where it is to be used in the solid state, in which it is most resistant to oxidization.

(C) EMULSION: Make a 1 per cent solution—or pseudo-solution—of sodium laurylsulfonate, or Duponol C, carefully checked to be sure that they are free from heavy metals, in distilled water. Then to 10 ml. of the above solution in a 250 ml. pyrex bottle add 5 ml. of chemically pure xylene. Xylene may be considered satisfactory if a 60 milligram per liter dithizone solution in the xylene does not lose more than 15 per cent of its strength in 7 days if the mixture is kept cool and in the dark.

Shake the xylene and laurylsulfonate—or Duponol C—until emulsified. Then add 10 ml. more of xylene and shake again until the whole

is emulsified. Continue adding xylene, shaking each time until emulsification is complete, until 90 ml. of xylene have been added; never add a greater volume of xylene at any one time than the volume of liquid already in the bottle. Incidentally, if no shaking machines are available and hand shaking must be employed, we recommend a 500 rather than a 250 ml. pyrex bottle. After all the xylene has been added, shake a further 5 to 10 minutes, or at least until the creamy emulsion does not separate on standing. This emulsion should then be tested. To do this add 1 ml. of emulsion to 100 ml. of slightly acid water and stir. After thorough mixing, the liquid should remain milky for from 15 to 20 seconds. If the solution takes longer than half a minute to clear, it may be just as efficient, or even more so, but it is less practical for field use. The attaining of just the right consistency in the emulsions still remains an empirical matter, probably best coped with by frequent sampling during the ultimate phase of shaking.

(D) HYDROCHLORIC ACID: Dilute concentrated acid with three volumes of double distilled water. If "chemically pure" acid is employed, it is usually pure enough not to need redistilling.

(E) ACETATE BUFFER: Dissolve 250 grams of hydrated sodium acetate and 11 ml. of acetic acid in 1 liter of distilled water. Purify by shaking with dithizone carbon tetrachloride solution until, on separating out, a mixed color (Huff 1948; Warren and Delavault, 1949) is obtained in the carbon tetrachloride layer, indicating the presence of an excess of uncombined dithizone.

(F) STANDARDS: Copper and zinc standards should always be available for testing dithizone solutions. We have found N/10 HCl solution containing 100 γ per ml. of zinc or copper to be a convenient concentration. For actual standardizing in the field, take 1 ml. of above standard and add 24 ml. of pure water giving a solution which contains 4 γ per ml. This is satisfactory for standardizing dithizone solutions but it must be used shortly after it has been prepared; we find that this weaker standard cannot be relied on for more than a few days.

BLANKS: Blanks should be run from time to time to test the purity of the reagents. In a laboratory, double distilled water is not always good enough and, on occasion, the task of obtaining pure water may assume major proportions. In British Columbia, it is often a relatively easy matter to obtain pure water, usually from melting snow and ice fields.

WATER TESTING

To 100 ml. of water in a pyrex open beaker, add 1.0 ml. of 3 N HCl, if a test for copper is to be made, or 1.0 ml. of acetate buffer, if a test for copper and zinc combined is desired. If the water is not reasonably "pure" and dilute, test for pH with a suitable testing paper to ensure that the water has been brought to pH 3.0 or 5.5 respectively. With exceptionally "pure" water, it may be sufficient if 0.2 ml. of HCl or buffer be added.

Next add a suitable amount of ammoniacal dithizone solution. This amount will naturally be found to vary from locality to locality and can only be learned by experiment and experience. In British Columbia, the most satisfactory amount has ranged from 0.2 to 1.0 ml., with the former figure best in most areas where there is a rainfall of 50 inches or more. This rainfall is not excessive but, because much of it falls as snow, it causes undue dilution during the runoff period.

Next add from 0.5 to 1.0 ml. of emulsion, stirring vigorously with a glass stirring rod. The thick creamy emulsion cannot be measured with precision but fortunately precision is not needed.

If there is no metal to be collected or very little, the milky liquid is greenish but, if there is sufficient to saturate the dithizone, the milky liquid becomes pinkish. It is useful to arrange a white background because weak colors may thereby be noted and contrasted more readily. In the field, where weather and insect life alike are apt to hinder calm and dispassionate observation, this precaution may be essential.

During a test, avoid exposing the beaker continuously to direct sunlight. If the milky liquid is green, no further test is needed, provided that the smallest practical amount of dithizone has been added in the first instance. If, however, the milky liquid turns pink, add more dithizone and more emulsion until the emulsion assumes a neutral color. The amount

of dithizone added represents a measure of the metal present in the water. For more accurate results repeat the test, taking 100 ml. samples of water and adding larger amounts of dithizone each test until a neutral gray is assumed by the emulsion.

In investigating any one area, an operator soon learns the normal metal content of the creeks in that area. He then adds a standard amount of the reagents, including enough dithizone to ensure its half saturation and provide a dark-green mixed color to the coalesced droplets of emulsion which rise to the top of the beaker. This corresponds to the metal content of the creek provided by decayed vegetation and scattered mineralization. The operator then tests as many samples of ground water as he can and looks for any variations in the color of the droplets of the emulsion as they rise to the surface. These droplets assume progressively, shades of greenish-gray, neutral gray, purplish-gray, very dark purple, dark purple, and purple as increasing amounts of heavy metals are encountered. Richer samples can best be tested by using less water and very rich ones by using the classical shaking method, diluting with pure water if that should become desirable. It must be repeated that this is a field method. Accurate results are not essential and the method, although well adapted to quantitative tests, finds its most ready application in a search for anomalous water. In many cases, it is quite unnecessary to attempt to translate the field data into numbers. Samples are taken until an anomaly is found. Then an attempt is made to discover the cause of the anomaly, in much the same manner as a prospector tries to find the source of some gold he has discovered in his pan.

Standardization

This should be done frequently because dithizone may start at any time to oxidize at a comparatively rapid rate. We have found it simplest to take from the standard solutions described above 8 γ of zinc or 20 γ of copper and place them in a ground-stoppered cylinder. The copper standard, being in a HCl solution, will already have the desired pH, about 3.0, but with zinc it will be necessary to add about 1 ml. of acetate buffer to get the desired pH, about 5.5–5.0. Measured amounts of the dithizone carbon tetrachloride solution are added and well shaken. This requires at least 2 minutes for copper but only about 20 seconds for zinc. The color is carefully noted and more dithizone carbon tetrachloride is added until the carbon tetrachloride layer assumes a neutral-gray color. Knowing the amount of dithizone carbon tetrachloride solution needed to produce neutral gray with 8 γ of zinc or 20 γ of copper, it is an easy matter to calculate the strength of the dithizone carbon tetrachloride solution. Because it is possible to extract virtually all this dithizone by shaking with an ammoniacal ($\frac{1}{100}$) solution, it is permissible, by using equal volumes of the ammoniacal and carbon tetrachloride solutions, to assume the same values for the ammoniacal as for the carbon tetrachloride solution.

However, it is possible to determine the strength of a dithizone ammoniacal solution with equal ease. The mixed colors produced by dithizone and metal dithizonates are about the same in xylene and in carbon tetrachloride; consequently, it is possible to place in a shaking cylinder a known amount of metal in solution, an equal amount of carbon tetrachloride (without dithizone), buffer if necessary, and note the colors obtained on shaking the mixture with increasing amounts of ammoniacal dithizone, measured with a pipette.

To test the visual ability of ourselves and the six prospectors with us we conducted the following demonstration. We took five 100 ml. beakers and placed in each 100 mls. of the purest water available. We then added from our stock solution, by means of a graduated pipette, to four of the beakers amounts of copper sufficient to make the copper content of the beakers corresponding to 3 γ, 6 γ, 10 γ, and 20 γ per liter respectively; the remaining beaker was left a blank. One ml. of 3 N HCl and 0.2 ml. of dithizone in the ammoniacal solution were then added to each, followed by approximately 1 ml. of the emulsion. All were stirred and the colors observed, both of the mass immediately after the stirring and of the emulsion at the surface when two minutes had been allowed for the emulsion to collect. All present were not only able to detect differences but they were also able to correlate the various concentrations

without trouble. A white background was found most helpful. This test was repeated several times in the field and except for an occasional anomaly never failed both to delight and instruct the members of our party.

Discussion of Method

It may be noted that, in discussing the emulsion technique, reference has been made to determinations of copper and of copper and zinc combined. It is possible to complex the copper with a thiosulphate solution and then determine the zinc, but owing to the great dilutions involved, it is not convenient to do this. Except near copper-bearing occurrences, where of course only copper determinations are needed, zinc is usually so predominant that we can neglect the error resulting from the differences between the mixed colors given by copper and zinc respectively with equivalent amounts of dithizone solution. Consequently, a negligible error is caused if the zinc content of the water is considered to correspond to the amount of zinc as shown by the acetate buffer reaction—which will include any copper present—minus the amount of copper which has been determined separately.

Other limitations of our technique should be noted. The emulsion and dithizone only reach free ions and as a result any metal absorbed by plankton will not be collected. Likewise, hydrochloric acid may liberate some metal adsorbed on mineral colloids, such as the clay minerals, and this may in occasional fortuitous circumstances result in some natural waters paradoxically appearing to contain more copper than copper and zinc combined, a feature which we found most disturbing in our earlier field work.

Furthermore, it must always be kept in mind that this emulsion technique, although extremely sensitive, is not very precise. In particular, the xylene emulsion will collect dithizone much more quickly than it will a metal complex. Furthermore, the collection of the metal complex is a surface phenomenon and is completed by the tiniest droplets, which of course have the greatest surface to volume ratio, and which consequently rise most slowly to join the thin layer of emulsion at the surface. As a consequence, when working near the limit of sensitivity for this method, it is possible to observe the tiniest droplets of emulsion with their metal content sufficient to tint the body of liquid a faint pink while the bigger droplets which have already risen to the surface may be a neutral gray, and even exhibit on less frequent occasions, a hint of greenish-gray. Thus it must be appreciated that there is a time element involved. However we have found that we may consider negligible the numerical errors which result from examining the surface layer before the last and tiniest particles of emulsion have risen to the surface to coalesce with the main layer of emulsion. What is essential is to follow a set procedure and to make this procedure as simple as possible.

Working with the extreme dilutions at which the emulsion technique is most useful, it is not possible to make a direct comparison with corresponding results obtained by the classical method. At such dilutions, the classical method gives only negative results.

Applications in the Field

This method has been put to use during two field seasons and has shown its value in exploration work. We are not at liberty to disclose all our results but we may state with propriety that we have discovered three mineralized showings as a direct result of water testing and biogeochemistry used in conjunction with one another.

In the most gratifying discovery of the three, zinc mineralization was traced to its source by means of a 3 γ per liter—3 parts per thousand million—anomaly originally detected over 2 miles away. Over 20 negative creeks were "cut out" by water testing and a high of 27 γ of zinc and 3 γ of copper were eventually recorded where the water emerged at the surface, about 100 yards from where mineral was eventually located.

Figure 1 illustrates the heavy metal content of numerous water samples taken in the vicinity of the Britannia Mine. The Britannia Mine is one of the two important copper mines operating in British Columbia. The area embraced by the sketch lies in a typically rugged section of the Coast Range of British Columbia. It lies about 20 miles due north of Vancouver and has

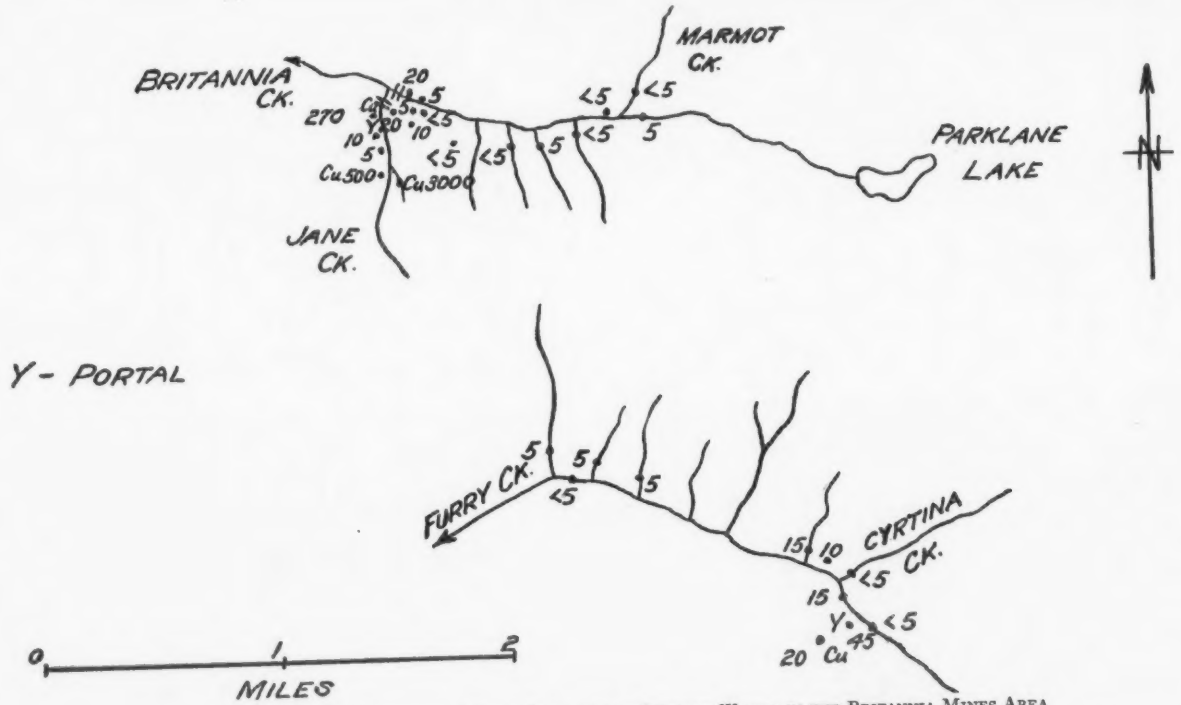

FIGURE 1.—SKETCH MAP SHOWING COPPER CONTENTS OF SURFACE WATER IN THE BRITANNIA MINES AREA

Numbers give the results of 29 copper determinations per thousand million parts of water, except for numbers below 20, where zinc content is included.

It should be noted how quickly even high concentrations may be diluted by a heavy run-off such as occurs in this area.

a rainfall ranging from 60 to 100 inches. The geology of the area has been well described by James (1929) and the figure has been prepared with the aid of a map presented in James' report. James writes of the area;

> "It is situated near the head of the most southerly of numerous irregular fiords, or inlets, which penetrate to the very heart of the Coast range. The mountains rise abruptly from these fiords to elevations of 6,000 and 7,000 feet. In the Britannia area, steep, uninterrupted slopes rise from sea-level to elevations of 4,500 feet within a horizontal distance of 7,000 feet. The mountains are cut into a series of ridges by deeply incised streams. The topographic relief even near the heads of the streams is usually more than 2,000 feet. The ridges between the streams are rounded, and vary in elevation between 4,500 and 5,500 feet".

To one not familiar with the area, it would seem reasonable to assume that outcrops would be abundant and prospecting no more difficult than might be expected in any rugged terrain. Outcrops are frequent but, as so often is the case, they are lacking in critical areas, and it is not surprising that, although copper was first discovered in the Howe Sound area in 1865, it was not until the 1890's that the Britannia showings were found (Hoffman, 1946).

The samples were taken during the last days of September 1950. This is the best possible time for this work because snow had disappeared from all but the highest mountains and heavy autumn rains had not yet commenced to unduly dilute the ground waters. A few springs and smaller creeks were still dried up.

The samples on Jane Creek containing 270, 3000, and 500 γ of copper carried also 2000, 9000, and 3600 γ of zinc respectively. The first two of these resulted from seepages from dumps and glory holes but the last owed nothing to man's efforts: it resulted from waters passing over a large zone of rubble containing ore caused by a slide on the mountain side. The 20 γ of copper found in Britannia Creek above Jane Creek were probably caused by a near-by mine dump. In the southern area, the water of the sample with 45 γ copper had less than 5 γ zinc and came from a trickle which had passed over an old dump with some seepage added from an abandoned tunnel. However, possibly the most relevant of all the samples was the 20 γ one taken close to the preceding one. This sample contained less than 5 γ of zinc, but the important point is that it was taken from a little spring above the portal and in ground completely undisturbed by man. This could have been found at any time.

It should be noted that Furry Creek less than a mile below its junction with Cyrtina Creek carried less than 5 γ of copper and zinc in spite of the fact that it was fed by at least one 15 γ copper and zinc stream below Cyrtina Creek, and that Furry Creek itself just above its junction with Cyrtina Creek carried 15 γ of copper and zinc combined.

Various observations of the authors point up the fact that frequently the major part of the water in a stream may sink through its gravel or sandy bottom and be replaced by seepages from its banks. Thus in a very short distance it is possible for a stream to lose its metal content, although the stream as a whole may exhibit no change in volume. Clearly it is always essential to remember physiography and drainage patterns in interpreting geochemical field results.

Furthermore, it is clear that in such areas as the humid slopes of the Coast range, mineralization will not provide much metal for natural waters and it would be easy to miss even important mineralization at some times of the year when the run off was particularly heavy. Certainly the major streams cannot be expected to exhibit measurable amounts of heavy metals except under unusually favorable circumstances. However it does seem probable that, by testing carefully all the minor streams and rivulets and all the springs, particularly at higher levels, any important mineralization can be discovered.

It seems clear that the amount of heavy metal in natural water is controlled more by the mineral assemblage and degree of weathering than it is by the absolute amounts of metal in the mineralization. A small amount of chalcopyrite if associated with any marcasite might yield much more copper to ground water than a rich bornite chalcopyrite ore.

Conclusions

The value of field analyses of surface waters as a tool in the never-ending search for ore has been established. The classical method described by Huff and others has its place where natural waters contain sufficient amounts of zinc and copper. The emulsion technique will

permit the range of search to be extended to areas where dilution of surface waters will not permit the use of the classical method.

The greater the sensitivity desired, the greater the difficulties involved in obtaining reagents and carrying out tests which will not be vitiated by salting.

Water testing is only another tool in the hands of a prospector or geologist and it should never be used without attempting to integrate it with any other tools which may be available.

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Manuscript Submitted to the Secretary of the Society, April 21, 1950

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 619-646, 5 FIGS., 2 PLS. JUNE 1951

GEOLOGY OF THE CEDAR HILLS, UTAH

BY STUART L. SCHOFF

ABSTRACT

The Cedar Hills, an area of about 320 square miles in central Utah between the northern end of the Wasatch Plateau and the southern end of the Wasatch Mountains, form the boundary zone between the Colorado plateaus and the Great Basin. The oldest exposed rocks are Carboniferous, but most of the area is underlain by Upper Cretaceous and Tertiary continental sediments many thousands of feet thick. The Indianola group, about 15,000 feet thick, consists principally of coarse conglomerates and sandstones indicative of near-by orogeny and contains a tongue of fossiliferous marine sandstones showing that it is of Colorado age. This group is overlain unconformably by thick fluviatile and lacustrine deposits ranging from Montana to Eocene in age (Price River, North Horn, Flagstaff, Colton, and Green River formations), which in turn are overlain unconformably by probably late Tertiary pyroclastics. The area was subjected to compressive orogenic disturbances in middle Cretaceous, upper Cretaceous (Montana), and probably middle Tertiary time, and to the normal faulting of the Basin-Range disturbance in late Tertiary time.

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Introduction

This paper presents data on the sedimentation and orogenic history of the Cedar Hills, central Utah, between middle Cretaceous and late Tertiary time. The area is 55 to 85 miles south of Salt Lake City (Fig. 1), has a maximum length of 30 miles, a maximum width of 15½ miles, covers about 320 square miles, and includes parts of Sanpete, Juab, and Utah counties and part of the Nebo Division of the Uinta National Forest.

The Cedar Hills are bounded on the east by the northern part of the Wasatch Plateau, where Spieker (1931; 1939; 1946) has found much that is new and significant on the nearshore marine and continental deposits of Upper Cretaceous and lower Tertiary age, and on the orogenic history of the region. West and north of the hills, the southern Wasatch Mountains are chiefly folded Paleozoic rocks and have been described by Eardley (1932; 1933; 1934). This study fills the gap between the areas covered by Spieker and Eardley.

Southwest of the Cedar Hills, the Gunnison Plateau continues the north-south trend of the Wasatch Mountains but is stratigraphically and structurally unlike them. Published ac-

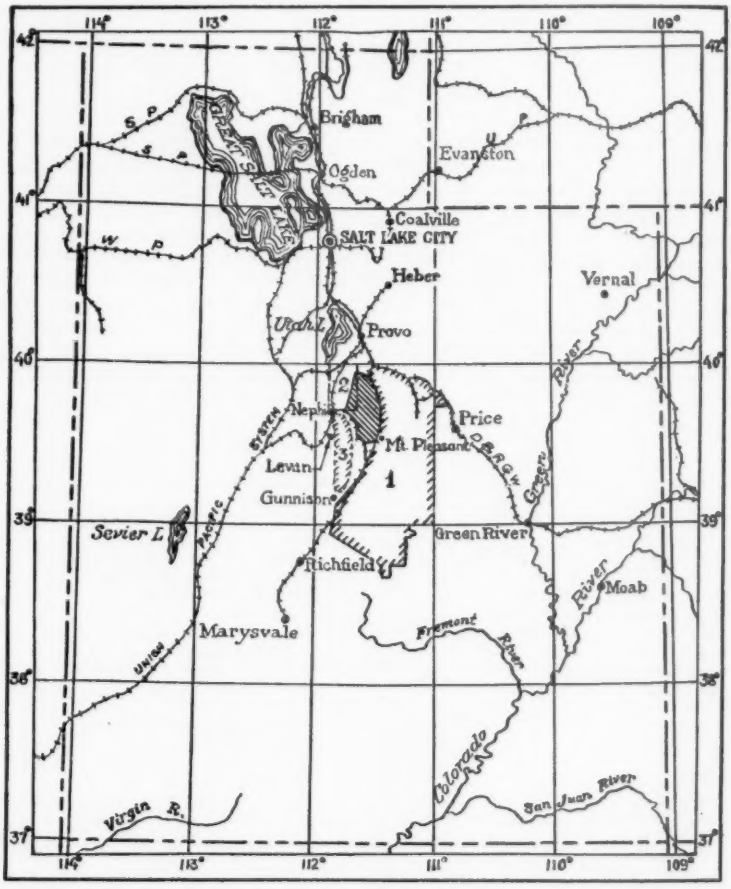

Figure 1.—Location Map

counts of the geology of this plateau are brief coal reports by Richardson (1906) and Clark, (1912).

South of the hills is Sanpete Valley, an alluvium-filled trough between the Wasatch and Gunnison plateaus. The principal towns in the mapped area—Mount Pleasant, Fairview, Moroni, and Fountain Green—are in this valley.

Previous Investigations

The Cedar Hills have received only brief treatment in earlier reports. The Wheeler Survey (1875) mapped the area as extrusive igneous rock and "Tertiary", and this precedent was followed on the geologic map of Utah (Butler *et al.*, 1920). Dutton (1880, p. 160) referred to the area as a "medley of low hills". Richardson (1907) included the southern end of the Cedar Hills on the geologic map accompanying a ground-water report on Sanpete and Sevier valleys but did not discuss it.

Field Work and Acknowledgments

Plane-table mapping by triangulation and traverse, on scales of 1 and 2 inches to the mile, was done in the summers of 1932 and 1936. Several of the standard iron corner posts set in 1916 by the General Land Office during a re-survey of T. 11 S., R. 3 E. were recovered and have been made the basis for adjusting the land net to the geologic map.

Much credit is due M. T. Sturgeon for his careful work throughout the 1936 field season, C. N. Schoff for assistance through the 1932 season, and D. T. Griggs, J. A. Kelly, and A. E. Schoff for assistance for shorter periods. E. M. Spieker, of The Ohio State University, spent several days in the field with the writer both in 1932 and 1936 and offered invaluable suggestions. The unpublished data he made available and his criticisms are especially appreciated. A. A. Baker and A. J. Eardley also criticized this paper, and the section on the fossils and age of the Indianola group was reviewed by J. B. Reeside, Jr.

The writer is indebted to the John A. Bownocker Fund of The Ohio State University for a grant applied on field work, and to the geology departments of that university and of Oberlin College for the use of field equipment. The Denver and Rio Grande Western Railroad furnished altitudes of points along the railroad. A. J. Eardley, of the University of Michigan, and L. H. Van Boskirk, of the United States Forest Service at Ephraim, Utah, furnished base maps. R. V. Kesling, of DePauw University, did part of the drafting.

Topography and Drainage

The Cedar Hills are a rugged, maturely dissected area. The lowest place in the area is Clinton station, in Thistle Valley, at an altitude of 5415 feet, and the highest is on the divide near the head of Nebo Creek at about 9400 feet. The northern part of the area, including all the perennial streams, is in the Great Salt Lake drainage basin. The southern part is in the portion of the Sevier Lake basin drained by the San Pitch River.

General Stratigraphy

The oldest rocks exposed in the Cedar Hills are Carboniferous. The older Paleozoic and Precambrian rocks exposed in the adjacent part of the Wasatch Mountains have been differentiated by Eardley (1933, p. 310) as follows:

Table 1.—Rocks Exposed in the Southern Wasatch Mountains

	Thickness (Feet)
Mississippian	
Brazer (?) and Madison limestones and dolomites	1650
Cambrian or post-Cambrian (pre-Mississippian) marine limestones and dolomites	2300
Cambrian	
Lower Middle Cambrian	
Ophir shale	200–300
Lower (?) Cambrian	
Tintic quartzite	900
Unconformity	
Algonkian red quartzites and shales	500–1000
Archean basal complex, schists, gneisses, granites, and pegmatities	?
Total	5550–6150

The rocks of the Cedar Hills range from Carboniferous to Tertiary and include sediments, extrusives, and metamorphics. The Carboniferous consists of fossiliferous marine lime-

stone and interbedded sandstone. The Jurassic is dominantly marine shale with some limestone, gypsum, and salt. Much of the Cretaceous is continental, but the included tongues of fossiliferous marine clastics are the key to the age of a very thick section of rocks. The Tertiary consists of continental sediments, except for the upper part, which is extrusive. The Cretaceous and early Tertiary (Eocene) formations aggregate between 21,000 and 28,000 feet.

Conglomerate is common in the formations from Upper Cretaceous to Eocene, and much of it is coarse, with boulders as much as 1 foot in diameter. Some of it in each formation contains pebbles of limestone derived from the

TABLE 2.—STRATIGRAPHIC COLUMN, CEDAR HILLS, UTAH

System	Series	Formation	Description	Thickness (Feet)
Quaternary	Pleistocene or Recent	Alluvium	Gravels, sands, and silts deposited by present streams	0–300
		Terrace Gravels	Gravels in low hills around margin of the area and above present floodplain levels	10+
Tertiary	Late Tertiary	Pyroclastic Rocks	Coarse sandstone with igneous pebbles; igneous-boulder conglomerate; massive white tuff; intruded near top by dark sill	1500
	Eocene	Green River	Shale, gray, black, and varicolored; limestone, thin-bedded to massive, partly fossiliferous; sandstones and conglomerates	4000?
		Colton	Red conglomerate, sandstones, and varicolored shale	580
	Paleocene	Flagstaff	Similar to Green River: shales, gray and white; limestone, sandstone, and some gray conglomerate	750
Cretaceous	Upper Cretaceous	North Horn	Sandstones, conglomerates, shales, and massive red limestones; lower part gray and brown, with fossil leaves; upper part dominantly red	1500–6700
		Price River	Conglomerate, massive, gray or red, with sandstone lenses; weathers to goblin forms where horizontal and to needles where tilted	134–1500
		Indianola Group	Conglomerates, sandstones, shales, and minor limestones; generally red; marine sandstones with Colorado invertebrates in upper part.	14,700±
Jurassic	Upper Jurassic	Arapien shale	Shale, gray drab to red, with thin interbedded limestones, gypsum, and salt. Only one small incomplete exposure in Cedar Hills	11,000± (Wasatch Mountains)
Carboniferous	Penn. and Permian	Oquirrh (?), etc.	Limestone, dense, dark-gray to black, interbedded with buff sandstones; only a part is exposed in Cedar Hills	10,000+ (Wasatch Mountains)

Carboniferous. The Jurassic, Cretaceous, and Tertiary shales are dominantly red, although other colors also occur. Limestone constitutes only a small percentage of the section above the Carboniferous, and above the Jurassic it is fresh-water limestone, in part algal.

Carboniferous System

Pennsylvanian and Permian

Distribution and lithology: The Carboniferous rocks of the area consist of interbedded fossiliferous limestone and sandstone in the Nebo Creek and Bennie Creek valleys. The limestone beds are 1 to 2 feet thick, gray and blue weathering to gray or buff, hard, fine-grained, and partly arenaceous. The sandstone beds are gray, red, brown or white, fine- to medium-grained, thin- to thick-bedded, and hard, siliceous, or quartzitic; locally they are cross-bedded.

Thickness: Exposures in the Cedar Hills represent only a small part of a very thick section better exposed outside the area, and a measurement of thickness therefore was not made. According to Eardley (1933, p. 322), the thickness of Pennsylvanian rocks in the southern Wasatch Mountains, including some possibly Permian strata, is 10,000 feet or more.

Fossils and age: Fossils from several localities include *Productus*, *Chonetes*, *Glossina* (?), *Neospirifer* cf. *triplicatus* (*dunbari*), a spiriferoid, Bryozoa, and fusulines (identifications by M. T. Sturgeon). These confirm in general the age reported by Eardley (1933, p. 322), in whose collections Girty found forms of Pottsville and younger Pennsylvanian age, but no Permian.

Nomenclature and correlation: The Carboniferous rocks are lithologically similar to the Oquirrh formation of the Oquirrh and Deep Creek Mountains, Utah (Gilluly, 1932), and part of them should probably be correlated with the Oquirrh. They are continuous with the intercalated series of Loughlin (1913, p. 447) and Eardley (1933, p. 322) in the southern Wasatch Mountains. Eardley suspected that Permian beds were included, but lacked paleontologic evidence. A. A. Baker (personal communication) later confirmed this suspicion by identifying Permian strata on the east side of Mount Nebo, and H. J. Bissell (personal communication) reports the Kirkman limestone, Diamond Creek sandstone, and Park City formation, all of Permian age, immediately north of Bennie Creek. With the Permian thus recognized both to the southwest and the north, it is probable that Permian strata are represented in the Cedar Hills along with the upper Oquirrh.

Triassic and Jurassic Systems

Woodside, Thaynes, and Ankareh Formations

Because of overlap of Upper Cretaceous and Tertiary formations, none of the Triassic and little of the Jurassic is exposed in the Cedar Hills, but rocks belonging to these systems crop out along Salt Creek, just west of the area. There Eardley (1933, p. 324–334) distinguished (1) the Woodside (?), consisting of limestones and red shales, 200 to 500 feet thick, thought to be Lower Triassic; (2) the Thaynes, consisting of buff, fossiliferous sandstones and arkose, 500 to 1000 feet thick, of Lower Triassic age; (3) the Ankareh, consisting of deep-red, arkosic sandstone, 1400 feet thick, assigned an "Upper Triassic or Jurassic" age; and (4) Upper Jurassic shale. According to Spieker (personal communication), Eardley's Ankareh is largely Navajo sandstone, or Nugget, of the Glen Canyon group. Only the Upper Jurassic shale is exposed in the area described in this paper.

Arapien Shale

The Upper Jurassic shale of Eardley (1933, p. 330–331) has been named the Arapien shale (Spieker, 1946, p. 123) from exposures in Arapien Valley, south of Mayfield, Utah. An elongate outcrop of the Arapien extends northward along the west side of the Gunnison Plateau, eastward around its north end, and northward along Salt Creek into the western part of the Cedar Hills.

The Arapien crops out near the mouth of Red Creek (secs. 9–10, T. 12 S., R. 2 E.), somewhat farther up Red Creek than Eardley (1934, map 11) shows it. It consists of red shale and yellowish, drab, or white fossiliferous limestone that is massive except where shattered.

At least two limestone layers make prominent ridges, one of them striking N. 20° W. and dipping 37° SW., and the other N. 40° E., 68° NW. These divergent strikes and dips are in keeping with the complex folding of the formation recognized elsewhere.

At most places, the thickness of the Arapien is difficult to measure because of folding. Eardley's highest estimate of the thickness in Salt Creek Canyon is 11,000 feet, and Spieker (1946, p. 125) states that the thickness in the Wasatch Plateau probably exceeds 10,000 feet. The small patch of Arapien shale in the Cedar Hills was not measured.

Fossils collected from the Red Creek exposure (SE¼ sec. 9, T. 12 S., R. 2 E.) by the writer were reported by Reeside to include: *Serpula* sp., *Ostrea* sp., *Cardinia* ? sp., *Trigonia montanaensis* Meek, *Camptonectes stygius* White, *Pleuromya* sp., *Pholadomya kingi* Meek, *Astarte* ? n. sp., and *Neritina* sp. This fauna, he stated, is of Jurassic age and about equal to that in the Carmel formation, of the San Rafael group. The Carmel is approximately equivalent to the Twin Creek formation of southwestern Wyoming and of the Wasatch Range at Salt Lake City, and to the Ellis formation of Montana (Baker *et al.*, 1936, p. 6–7).

Cretaceous System

Indianola Group

Definition and type locality: The oldest Cretaceous rocks recognized in central Utah are a thick series of conglomerates, sandstones, and shales named the Indianola group (Spieker, 1946, p. 127). At the type locality north of Indianola, Utah, this group is not clearly divisible, but 40 to 60 miles to the south, in the Manti-Salina area, four units are distinguished. The Indianola group also has been recognized on both the east and the west sides of the Gunnison Plateau.

Distribution: In the Cedar Hills, undifferentiated Indianola underlies most of the Pole Creek drainage basin and the northwestern side of the Hop Creek basin. This outcrop connects those in the Gunnison Plateau and the northern part of the Wasatch Plateau except for short gaps.

Lithology: The Indianola consists of alternating conglomerates, sandstones, shales, and a few thin beds of fresh-water limestone. Individual beds range from 1 foot or less to more than 50 feet thick; some can be traced as much as 1 mile, but others disappear in short distances. The upper and lower parts are continental and dominantly red. Between them is a tongue consisting chiefly of gray and white, fossiliferous marine sandstone.

Pebbles in the conglomerates range from a fraction of an inch to 6 inches in diameter and consist of quartzite of various colors and dark-gray to black Pennsylvanian limestone. The limestone pebbles are abundant in some beds but scarce in others. A conglomerate near the top is made up principally of algal nodules. The sandstones range from fine to conglomeratic, from thin-bedded to massive, and are white, gray, yellow, yellowish-gray, pink, red, and brown. In the lower part of the section they are highly calcareous.

Generalized Section of Indianola Group, Pole and Hop Creeks, Cedar Hills, Utah

	Thickness (Feet)
Price River formation	
Angular unconformity	
Indianola group	
(3) Alternating conglomerates and sandstones; fossil leaves in some sandstone beds; completely cut out by overlap of Price River formation at northeastern end of Hop Creek Ridge; maximum exposed thickness	2800
(2) Marine fossiliferous sandstones and interbedded conglomerates; sandstones range from fine to conglomeratic, from thin- to thick-bedded, and in part are cross-bedded; some of the fine, fossiliferous sandstones are friable, but the coarse sandstones are well cemented; exposed thickness	2530
(1) Conglomerates, sandstones, red shales, and gray to white limestones. The limestones are fresh-water and algal types, and in lower 7000 feet are thin, discontinuous, and drab, or mottled red and yellow. About 2300 feet below the top, a 20-foot, white-weathering, dense, drab, irregularly bedded limestone with irregularly rounded nodules and jasper is prominently exposed for about 1 mile along the southeast side of the Middle Fork of Pole Creek. A second and thinner limestone is less well exposed several hundred feet higher in the section. About 4400 feet of this unit is concealed under a cap of pyroclastic rocks on the divide between the Left and Middle Forks of Pole Creek	9350
Total	14,680

THICKNESS: As measured by stadia, the Indianola of the Cedar Hills is about 14,700 feet thick. The top is fairly well exposed, but the base is concealed by overlap of younger formations, and the maximum thickness, therefore, may be greater. Further uncertainty is due to the concealment of about 4400 feet of beds in unit 1 by pyroclastics between the Left and Middle forks of Pole Creek. However, the lithologic and structural similarity on the east and west sides of this covered area supports the assumption that the section under cover is continuous. Isoclinal folds, or faults, might possibly repeat part of the section under the pyroclastic cover, or elsewhere in the outcrop area, but evidence of such structures is lacking, and in terms of regional habit they seem most unlikely. If the estimate given here is correct, the Cedar Hills contain one of the thickest sections of Upper Cretaceous rocks in the Cordilleran region, and the area probably lay close to the axis of an Upper Cretaceous geosyncline.

RELATION TO OLDER FORMATIONS: The base of the Indianola group is concealed in the Cedar Hills by the unconformably overlying Price River formation and the pyroclastic rocks on the Pole Creek-Red Creek divide, and by the covered interval along the lower part of Salt Creek valley. Probably it is approximately conformable to the underlying Arapien shale, as it is on the west side of the Gunnison Plateau, near Levan, Utah.

FOSSILS AND AGE: Collections of pelecypod and gastropod molds and casts from fine-grained sandstones of the lower part of unit 2 were obtained at several localities along the Right Fork of Pole Creek and Rolley Canyon, and were found by Reeside to include *Pteria gastrodes* Meek, *Ostrea* sp. (also *O.* a small form), *Inoceramus* sp., *Cuspidaria* sp., *Liopistha meeki* White, *Cardium pauperculum* Meek, *Legumen* aff. *L. planulatum* Conrad, *Tapes* aff. *T. cyprimeriformis* Stanton, *Dosiniopsis oribculata* Hall and Meek, *Veniella goniophora* Meek, *Mactra utahensis* Meek, *Turbonilla coalvillensis* Meek, and *Turritella* (?) sp. Regarding them he wrote:

"This fauna occurs at many horizons in the Colorado group in sandy beds. It is apparently a facies fauna rather than a limited chronologic assemblage. The best that can be said is that it is of Colorado age, probably lower."

Two collections of fossils representing the upper part of unit 2 at the northeastern end of Hop Creek Ridge (sec. 17, T. 12 S., R. 3 E.) were found by Reeside to include *Ostrea prudentia* White, *Liopistha concentrica* Stanton, *Cyrena* ? sp., *Tellina* ? *isonema* Meek, *Mactra utahensis* Meek, and *Corbula nematophora* Meek. This assemblage, he said, indicates early Colorado age. Two collections also from Hop Creek Ridge, but about 3 miles to the southwest along the strike of the beds, contained other species and indicated late Colorado age. One of hese collections (from sec. 26, T. 12 S., R. 2 E.) contained *Membranipora* sp., *Ostrea* sp., *Cardium pauperculum* Meek, *Aphrodina* ? sp., *Tapes cyprimeriformis* Stanton, *Legumen* ? sp., *Tellina* ? *subalata* Meek, *Mactra arenaria* Meek, *Fusus* ? sp., *Baculites* sp., and an ammonite (?) fragment, and was designated by Reeside as a late Colorado fauna, comparable to that near the top of the Colorado at Coalville, Utah. The other, from sec. 24, T. 12 S., R. 2 E., contained the same *Tellina* and *Mactra* and a *Membranipora*, and in addition *Serpula* sp., *Inoceramus deformis* Meek, *Ostrea* cf. *O. anomioides* (Meek), *Cardium curtum* Meek and Hayden, *Capulus* ? n. sp., and *Architectonica n. sp.* Reeside considered this a lower Niobrara fauna.

Field evidence suggests that all four collections from Hop Creek Ridge came from a zone a few hundred feet thick, but it is impossible to state positively that the "late Colorado" collections came from beds above the "early Colorado" collections. The paleontologic determinations involve *Ostrea prudentia*, regarding which Reeside wrote (personal communication).

"Occasionally a form that has been beyond question appears to turn up in a horizon beyond its assumed range. We have either to accept it as an error in collecting or a more or less accidental parallelism in form on the part of some other species. In your collections I have identified *Ostrea prudentia* White in lots 16198 and 17074 (from the northeastern end of Hop Creek Ridge). All previous records of this species are from the basal part of the Colorado group and I therefore originally assigned the lots to the lower Colorado. If, however, the two lots are to be associated stratigraphically with lots

17265 and 17266 (from secs. 26 and 24), I would say that the whole fauna represented in the four lots is of Niobrara age, and that previous ideas as to the range of *O. prudentia* must be revised." (Material in parentheses added by author).

Therefore, further collecting and field study of the stratigraphic relations of the two localities is desirable before the significance of *O. prudentia* may be known.

Unit 3 of the Indianola has yielded only a few fossil leaves. Material from the lower part of this unit on Hop Creek Ridge (near SE. cor. sec. 33, T. 12 S., R. 2 E.) was identified as *Dewalquea* cf. *D. pulchella* Knowlton, approximately of Frontier age, by Roland W. Brown. He considered a fragment of a dicotyledonous leaf from the upper part of the unit as not determinable. As the Frontier belongs to the Colorado, this meager evidence corroborates the evidence of the more abundant invertebrates.

Unit 1 of the Indianola, aggregating 9350 feet of sediments, has yielded no fossils and is assigned to the Colorado because no break is evident between it and the overlying fossiliferous strata. Possibly part or all of it is Lower Cretaceous, but if so it is an exception, for Lower Cretaceous rocks have not been recognized elsewhere in central and eastern Utah. A local development of Lower Cretaceous sediments in the Cedar Hills area, however, might be attributed to the accumulation of fluvial gravels on the piedmont slopes of newly formed mountains before deposition of Upper Cretaceous rocks began in the basin farther east.

Possibly, also, some of unit 1 is equivalent to the Morrison formation. The steeply dipping sandstones, conglomerates, and red shales along the northwest side of the Left Fork of Pole Creek, and the beds concealed by the pyroclastics between the Left and Middle forks, might be the Morrison. A plausible upper boundary was not found, hence the thickness is uncertain, but it would be about 5000 feet if all the concealed beds were included. This thickness is more than previously recognized in Colorado, New Mexico, and Utah (Baker *et al.*, 1936, p. 9) except for the north flank of the Uinta Mountains at Coalville, Utah, where Eardley (personal communication) reports 3000 feet of Morrison (?). Spieker (1946, p. 125–126) discusses the Morrison from a regional point of view, concludes that the formation should be present in central Utah, and assigns to it 1300 feet of strata near Salina and 1800 feet near Thistle.

Fossils from the Indianola group of the Cedar Hills clearly indicate Colorado age but represent a zone only 2500 to 3000 feet thick, which is about 20 per cent of the total thickness. The lowest fossils were found more than 9000 feet above the base, and the highest were about 2300 feet below the top. Assignment of the whole group to the Colorado is based on the available fossils and the apparent continuity of the strata but obviously may require revision.

CORRELATION: The Indianola group in the Cedar Hills is the southwestward extension of the Indianola from its type locality in the western part of the Wasatch Plateau. The two outcrops are only 5 miles apart, and similarities in stratigraphic position and lithology leave little doubt of their equivalence. These similarities are also evident in the Gunnison Plateau, to the southwest. The only fossil thus far found in the Indianola of the type locality, *Corbula nematophora* Meek, occurs also in the Indianola of the Cedar Hills. It is suggestive but not conclusive.

The sediments of the Indianola become finer to the east, and between the western and eastern sides of the Wasatch Plateau they grade

PLATE 1.—PRICE RIVER AND ADJACENT FORMATIONS

FIGURE 1.—MOUTH OF ROUGH CANYON

(a) Unit 3, Indianola group (b) conglomerate needles and (c) soft sandstone in Price River; (d) lower North Horn.

FIGURE 2.—NORTHEASTERN END OF HOP CREEK RIDGE

Indianola group (a) overlapped by Price River (b); (c) lower North Horn.

FIGURE 3.—PSEUDO-UNCONFORMITY AT PRICE RIVER—NORTH HORN CONTACT

Upper Hop Creek Valley.

FIGURE 1

FIGURE 2

FIGURE 3

PRICE RIVER AND ADJACENT FORMATIONS

into the sandstones and shales of the Mancos formation (Spieker, oral communication). Along the west side of the plateau, equivalent beds totaling 7905 feet contain much marine material (Spieker and Reeside, 1926, p. 431–432). According to Spieker (1946, p. 127–128), this section is divisible into four formations, the Sanpete (lower Colorado), Allen Valley shale (middle Colorado), Funk Valley (Colorado), and Sixmile Canyon (Colorado). He considers the Funk Valley as probably equivalent to the marine zone (unit 2) of the Indianola of the Cedar Hills.

In the Kaiparowits region, the approximate equivalents of the Indianola group are the Tropic shale and Straight Cliffs sandstone (Gregory and Moore, 1931, p. 95–104).

CONDITIONS OF ORIGIN: While the Cretaceous marine embayment was still far to the east, streams draining new western mountains spread gravel and sand on the piedmont, making unit 1 of the Indianola in the Cedar Hills. When the sea had invaded the Cedar Hills, the fine-grained, fossiliferous sandstones of unit 2 were deposited. Streams draining the western highlands were still vigorous, and they brought gravel that was deposited with the sand in the shore zone. As the sea began to retreat eastward, the massive conglomerate now at the crest of Hop Creek Ridge (base of unit 3) was deposited, perhaps on the beach. In the overlying beds, leaves preserved in sandstone indicate a humid climate and suggest that the red of some associated shales is doubtless due to derivation from older red beds, such as the Arapien shale. The absence of coal or carbonaceous shale indicates good drainage.

Price River Formation

TYPE LOCALITY: Spieker and Reeside (1925, p. 445) named the Price River formation from Price River Canyon northwest of Castelgate, Utah. Southward from the type locality through the eastern part of the Wasatch Plateau the formation consists of a massive, cliff-forming sandstone, which Clark (1928, p. 20) named the Castlegate member, and overlying sandstones and shales. Westward it becomes coarser, and in the northwestern part of the Wasatch Plateau it is massive conglomerate (Spieker, 1946, p. 131).

DISTRIBUTION: The conglomeratic phase of the Price River crops out west and north of Red Creek and extends eastward toward the Left Fork of Pole Creek and northward to Nebo Creek. It caps the Nebo Creek-Salt Creek divide as far north as the head of Black Canyon and Holman Creek. Another outcrop extends from the mouth of Hop Creek northeastward into the headwaters of Spencer Creek, where it disappears under Tertiary pyroclastic rocks.

LITHOLOGY: Along Red Creek, the Price River formation is massive red conglomerate with a red and gray matrix. Boulders up to 1 foot in diameter are mostly well-rounded quartzite from Cambrian and older formations. Pebbles of Pennsylvanian limestone are common. The lenticular sandstones are dark red, weather to tan or yellow, and constitute only 10 to 20 per cent of the total thickness. Northward and eastward in the Cedar Hills, the proportion of sandstone increases so that in the Nebo Creek basin and along the Left Fork of Pole Creek the Price River formation is difficult to distinguish from the overlying North Horn formation. Still farther north the Price River-North Horn part of the section becomes so conglomeratic that A. A. Baker (personal communication) finds it impracticable to separate them in the Strawberry Valley area. Besides conglomerate and sandstone, the Price River along Red Creek has a little shale containing gastropods.

Along Hop Creek, the Price River is dominantly gray. At the mouth of Rough Canyon the formation is nearly all conglomerate (Pl. 1, fig. 1):

SECTION OF PRICE RIVER FORMATION

At the mouth of Rough Canyon (NW¼ sec. 30, T. 12 S., R. 3 E.)

	Thickness (Feet)
North Horn formation	
Price River formation	
(3) Conglomerate, with pebbles of quartzite, sandstone, and dark-gray limestone up to 8 inches in diameter, average 3 or 4 inches; coarse, brown and gray matrix. Interbedded thin, buff and pink sandstones	75
(2) Sandstones, white, massive, fine-grained, friable	175
(1) Conglomerate, massive, coarse; pebbles and cobbles averaging 5 inches, in light-colored matrix of angular grains; lower 500 feet somewhat reddish because of red and pink quartzite pebbles and iron oxide stains; conglomerates somewhat blackened by lichens; contains a reddish limestone layer	1250

Disconformity

Indianola formation
Total..... 1500

The proportion of sandstone increases along the strike in both directions from the above locality. At the head of Hop Creek, 2¼ miles northeast, nearly half of a 1200-foot section is very coarse, conglomeratic, gray and brown sandstone, in part cross-bedded.

THICKNESS: Thicknesses of the Price River formation differ greatly. About half a mile southeast of the mouth of Black Canyon (sec. 24, T. 11 S., R. 2 E.) only 134 feet was assigned to the formation, but near the head of Red Creek it is about 1000 feet thick. To the east, in the northwestern Wasatch Plateau, it ranges from 1000 to 2000 feet thick (Spieker, 1946, p. 132).

RELATION TO OLDER FORMATIONS: Along the southeastern base of Hop Creek Ridge, both the Price River formation and the Indianola group strike 50° to 60° NE. and dip 75° or more SE. At some places the two are approximately conformable, but at the northeastern end of Hop Creek Ridge the strike of the ridge-making conglomerate at the base of unit 3 of the Indianola changes from N. 60° E. to N. 80–90° E., and the conglomerate terminates against the Price River formation, which maintains its strike of about N. 50° E. (Pl. 1, fig. 2). The Indianola is vertical or slightly overturned, but the Price River dips 75° SE. In short, there is an angular unconformity of 40° or 50° in strike between the two formations, and unit 3 of the Indianola is completely cut out by overlap. Probably unit 2 is similarly cut out to the northeast in the headwaters of Spencer Creek.

At several places nearly horizontal or slightly tilted Price River beds overlie steeply dipping older formations. They are on steeply tilted Indianola along the northwest side of the Left Fork of Pole Creek, Ankareh at Cow Hollow (sec. 17, T. 12 S., R. 2 E.), Arapien shale along the west side of Red Creek, and moderately tilted Carboniferous along the upper part of Nebo Creek.

FOSSILS AND AGE: Direct evidence of the age of the Price River formation in the Cedar Hills is negligible. In a small collection of fossils from the Price River formation west of the Left Fork of Pole Creek, Reeside identified *Hyalina* ? cf. *H. evansi* Meek and Hayden, *Viviparus* cf. *V. prudentia* White, and *V. leai* Meek and Hayden. He stated that these are fresh-water forms described from Cretaceous rocks. The fact that the Price River overlies unconformably the Indianola group, which contains Colorado fossils, indicates that it probably is post-Colorado, and regional relationships point to late Montana.

Similar lithology and stratigraphic position suggest that the Price River of the Cedar Hills is equivalent to the conglomeratic Price River in the western Wasatch Plateau. The latter grades eastward into the sandstones and shales of Price Canyon and the eastern Wasatch Plateau, which are probably late Montana (Spieker, 1946, p. 131).

CORRELATION: The Price River is approximately equivalent to part of the Lewis shale and the Pictured Cliff, Fruitland, Kirtland, and McDermott formations of the San Juan Basin of Colorado and New Mexico (Reeside, 1924, p. 18, 19, 24, 27); the Fox Hills of Colorado and the Dakotas (Reeside, 1933, Pl. 8; O'Harra and Connolly, 1926, p. 41); and part of the Lewis shale of Wyoming (Reeside, 1933, Pl. 8).

CONDITIONS OF ORIGIN: The Price River is a widespread formation that grades from coarse continental clastics eastward into marine sediments. After the folding and thrusting of the southern Wasatch Mountains in the Montana (early Laramide) orogeny and the truncation of the fold in Hop Creek Ridge, eastward-flowing streams dumped coarse gravel and sand over the area of the Cedar Hills and northwestern Wasatch Plateau and carried finer-grained materials into the sea beyond, to be deposited as the Castlegate sandstone and upper Price River formation. The Castlegate continues eastward across Utah as a sandstone tongue in the Mancos shale (Spieker and Reeside, 1925).

CRETACEOUS AND TERTIARY SYSTEMS

North Horn Formation

DEFINITION AND TYPE LOCALITY: The North Horn formation is defined by Spieker (1946, p. 132–135) as variegated fluviatile and lacustrine beds containing reptile and mammal remains, between the Price River formation and the Flagstaff limestone. It is named for North Horn Mountain in the Wasatch Plateau. Prior to the discovery in it of dinosaur remains and Pale-

ocene mammals, it was classified as the lowest member of the Wasatch formation (Spieker, 1931, p. 45–46).

Distribution: In the Cedar Hills, the North Horn underlies most of the Hop Creek drainage basin; large areas in the drainage basins of Nebo and Bennie Creeks; small areas in canyons west of Payson ranger station; and a narrow strip around the head of the Middle Fork of Pole Creek.

Lithology: The North Horn formation contains many lithologic types. The sandstones and conglomerates range from fine to coarse and from gray and white through brown to pink and red. The shales generally are red, but some are gray. The limestones are thick or thin, gray, drab, or pink, and dense and hard; some are algal, some sandy. Few beds can be traced far laterally, and vertical graduation is the rule. In many exposures the formation is red, but the lower 1500 to 3900 feet is mostly gray and brown.

In the lower unit, which has a maximum thickness of about 3900 feet and is recognized only along Hop Creek, the gray and brownish-gray conglomerates contain pebbles of brown, pink, and gray sandstone and dark-gray limestone. The sandstones are gray, buff, brown, and pink, range from very fine to coarse and from thin-bedded to massive, and in part are cross-bedded. The few limestones are thin. A 430-foot interval beginning about 1580 feet above the base probably consists chiefly of shale, as indicated by a few small exposures of gray, buff, yellow, chocolate-brown, and red shale.

The upper unit of the North Horn is about 2700 feet thick in the Hop Creek basin. At its base is a massive, white, crag-forming conglomerate 100 to 200 feet thick that crops out on the southeast side of the basin. The overlying beds are red and gray conglomerates and sandstones interbedded with red shales. They are especially well exposed in the head of a tributary to Spencer Creek about 1 mile northwest of Mount Baldy (SE¼ sec. 16, T. 12 S., R. 3 E.), where the conglomerates range from 3 to 15 feet thick, from fine to coarse, and from red to gray. Cobbles and boulders from 3 inches to 1 foot in diameter consist largely of well-rounded white, purple, and red quartzite, but some of the smaller ones are sandstone. The sandstone beds are 6 to 8 feet thick, white, gray, yellow, red, and pink and range from fine to coarse. The red, yellow, and gray shales are 3 to 12 feet thick and in part are sandy. The interbedded sandstones and conglomerates weather in fantastic forms.

The upper unit appears to be the only part of the formation represented in the Nebo and Bennie Creek areas unless part of the beds along Wheatgrass Creek (secs. 21 and 28, T. 11 S., R. 3 E.) represent the lower unit. In the Nebo and Bennie Creek areas the congloer-so ates, sandstones, and shales are similar to them of Hop Creek, but fresh-water limestones with algal nodules are thicker and more abundant.

THICKNESS: The North Horn ranges from 4200 to 6700 feet in thickness in the Hop Creek basin, partly due to northeastward thinning of the lower unit as a result of original differences in sedimentation. Near the mouth of Rough Canyon, the lower unit is 3520 feet thick, and about 1½ miles to the northeast it is 3950 feet, but 3 miles farther northeast, at the head of the basin, it is only 1650 feet.

Elsewhere in the Cedar Hills, a complete section of the formation was not measured. Probably at least 1000 feet, but not over 2000 feet, of the upper unit is exposed in Golden Ridge, in the Nebo Creek basin. At the bend in Wheatgrass Creek (secs. 21 and 28, T. 11 S., R. 3 E.) 3100 feet of North Horn is exposed, but, as the base is concealed by pyroclastic rocks, this is not the complete thickness for this locality, and the thickening southeastward from Golden Ridge to Wheatgrass Creek therefore exceeds 1100 feet.

The North Horn formation of Hop Creek and Wheatgrass Creek is exceptionally thick, and these localities doubtless were near the axis of the basin of North Horn sedimentation. The line connecting them—a line striking northeast—may approximate that axis.

RELATION TO OLDER FORMATIONS: The contact of North Horn on Price River has been mapped at the top of the thick, massive conglomerate where there is one, as in Hop Creek. In the upper part of the Nebo Creek basin and along the Left Fork of Pole Creek, the Price River is sandier, and a less satisfactory boundary has been drawn above the more prominent conglomerates. Owing to lithologic gradation

both vertically and horizontally, this boundary is not a true stratigraphic horizon.

A relationship suggesting angular unconformity between the North Horn and Price River formations is evident along the ridge between Hop Creek and Rough Canyon. As seen

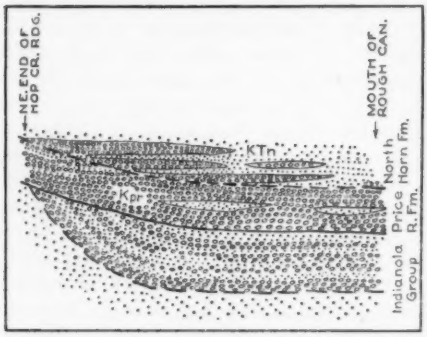

FIGURE 2.—PSEUDO-UNCONFORMITY EXPLAINED AS INTERFINGERING OF SANDSTONES AND CONGLOMERATES

from a couple of miles farther south, white outcrops of massive Price River conglomerate appear to truncate adjacent beds of the North Horn, the strike of which is indicated by lines of trees (Pl. 1, fig. 3). The implication is that the younger beds are on the northwest and the older on the southeast, and therefore that the formations are wrongly identified, inasmuch as an older formation (Price River) cannot truncate a younger one (North Horn).

Regional relationships show that the apparent truncation and unconformity are deceptive. Southeastward from the southern Wasatch Mountains across Rough Canyon, the succession of formations is from Upper Jurassic to Tertiary—from older to younger, as is normal on the flanks of folded mountains. In this succession, younger beds are on the southeast, older ones on the northwest.

At close range, the truncation seems to vanish in a maze of North Horn sandstones and conglomerates, each striking a little differently from its neighbor, yet not greatly different from the Price River. Hence the "truncation" seems to be a cross section of interfingering beds (Fig. 2), as suggested by Reeside (oral communication). It is sedimentary transition on display—a hesitant transition in which sand tongues were covered by gravel tongues continuous with the main body of the Price River. The line of apparent truncation connects the bases of the conglomerate tongues.

Where the North Horn formation rests on the Carboniferous rocks, as along Holman, Nebo, and Bennie creeks, there is an angular unconformity of 20°–25°. Also, along Nebo Creek the altitude of the contact rises and falls considerably without relation to the dip of the beds, suggesting relief of several hundred feet on the unconformity.

FOSSILS, AGE, AND CORRELATION: The fossils from the North Horn in the Cedar Hills are poorly preserved leaves and invertebrates obtained from sandstone beds. In two small collections from the lower unit of the formation on the west side of Hop Creek (sec. 30, T. 12 S., R. 3 E.), R. W. Brown found *Platanus* cf. *P. guillelmae* Göppert, probably of Paleocene age, and indeterminable fragments of palm and dicotyledonous leaves. He reported a collection from the upper unit that was obtained on the east side of the left fork of Beaver Dam Creek (sec. 6, T. 11 S., R. 3 E.) as fragments of dicotyledonous leaves.

Two collections of invertebrate fossils were obtained from the North Horn along Beaver Dam Creek, one at the forks and the other on the east side of the left fork (sec. 6, T. 11 S., R. 3 E.). In them, Reeside found *Goniobasis nodulifera* Meek, *G. tenera* Hall, *G.* sp. heavily coated with an algal deposit, and *Helix riparia* White, and *Helix* cf. *H. peripheria* White. He wrote:

"This fauna has been called Eocene in past years and seems to range as high as that. However, it probably occurs also in the Paleocene and could be in the upper North Horn. It is not known in Cretaceous beds."

Poorly preserved fossils were found also in beds along Wheatgrass Creek and an unnamed creek in secs. 24–25, T. 11 S., R. 2 E.

The North Horn formation of the Cedar Hills is of Paleocene and perhaps early Eocene age, and it may also include Cretaceous strata. It occupies about the same stratigraphic position as, and is lithologically similar to, the North Horn formation of the Wasatch Plateau, but it cannot be traced across the intervening area. In the plateau, dinosaur remains have proved the Cretaceous age of the lower part,

which grades upward into beds containing two Paleocene mammalian faunas (Gilmore, 1938, p. 1; Gazin, 1938, p. 271; 1939, p. 273–274; 1940, p. 8). Whether the North Horn formation of the Cedar Hills includes Cretaceous strata has not been proved, but the transition from the Price River formation, of Cretaceous age, into the North Horn points to the probability. The 1000 feet of barren beds between the Price River and the horizon of the first leaf of "probably Paleocene" age might well be Cretaceous.

Tertiary System

Flagstaff Formation

Definition and type locality: The Flagstaff formation was named by Spieker and Reeside (1925, p. 448–449) from Flagstaff Peak in the southern part of the Wasatch Plateau, where it consists chiefly of fresh-water limestones described as "dense, fine-grained . . . cream to tan in color." Beds of blue-gray limestone with casts of gastropods and pelecypods and other beds consisting largely of fresh-water ostracods are also common. Formerly classified as the middle member of the Wasatch formation, the Flagstaff is now ranked as a formation (Spieker, 1946, p. 135; Wilmarth, 1938, p. 737).

Distribution: In the Cedar Hills, a succession of light-gray and white strata including some limestone overlies the North Horn, cropping out in Water Hollow and southwestern Big Hollow and along creeks in the northern part of the area. These beds occupy the stratigraphic position of the Flagstaff formation of the Wasatch Plateau but contain more clastic material.

Lithology: The Flagstaff in the northern Cedar Hills is thick, cliff-making limestone which Eardley (1932, p. 403–405) described as "oöidal" because nodules, probably of algal origin, are abundant in it.

Farther south, at the head of Beaver Dam Creek, the Flagstaff is represented by three or four massive limestones separated by red beds. The limestones are gray to pink, weather white, and contain algal nodules and fresh-water gastropods; some of the latter are heavily coated with algal growth. The intervening red sandy shales, sandstones, and thin conglomerates are similar to parts of the North Horn formation.

Still farther south, in the Water Hollow area, the basal bed of the Flagstaff is a yellowish-buff, massively bedded, fresh-water limestone about 50 feet thick, containing algal nodules and pink and white quartzite pebbles. It crops out along the Hop Creek-Water Hollow divide. The higher limestones are of several types—some algal, irregularly bedded, white; some thin-bedded, platy; and some oölitic. Locally, some of the limestone is silicified. Although limestone beds occur in many exposures in Water Hollow, the dominant lithology there is clastic. There are gray sandstones, medium- to coarse-grained, thin-bedded to massive, in part calcareous; gray shales; and gray conglomerates containing the same kinds of pebbles as the conglomerates of older formations.

Thickness: The exposed thickness of the Flagstaff along Beaver Dam Creek is about 300 feet, but an unknown thickness has been eroded. The thickness in Water Hollow is greater, perhaps 750 feet; the exact amount is uncertain because at least one fault cuts the beds.

Relation to older formations: The Flagstaff is gray to white, contrasting with the red in adjacent formations. The contact with the underlying North Horn formation is at the base of the lowest more or less continuous light-colored bed, and probably is gradational, doubtless marking a change of environment. The interbedded light-colored limestones and red clastic deposits at the head of Beaver Dam Creek probably represent intertonguing of the Flagstaff and North Horn formations.

Persistent light-colored beds do not crop out in Golden Ridge (southwestern T. 11 S., R. 3 E.) where they might be expected to appear, and the ridge, therefore, is mapped entirely as North Horn. This interpretation is subject to revision if the Flagstaff wedges out so that North Horn beds pass directly into the Colton formation, or if the red limestones in Golden Ridge, here included in the North Horn, should be Flagstaff. If either of these possibilities is correct, the thickness of 1000 to 2000 feet for the North Horn in Golden Ridge is too great.

Eardley (1932, p. 399) reports Flagstaff limestone resting on truncated Carboniferous rocks about 3 miles west of the area of this report, at Tinnie Flat, a locality thought by him to have been near the foot of the mountain range of Flagstaff time.

FOSSILS AND AGE: Fossil collections from the southwestern part of Big Hollow include *Sphaerium* sp., *Limnaea* or *Succinea* sp., *Planorbis* sp., identified by Reeside, who reported that they are a fresh-water assemblage once assigned to the Eocene but now of doubtful age.

The Flagstaff formation of the Cedar Hills is thought to correlate with the Flagstaff of the Wasatch plateau where, according to Spieker (oral communication), species in the upper part may be of Wasatch age and those in the lower part pre-Wasatch. The Flagstaff, then, is of upper Paleocene and lower Eocene age.

CONDITIONS OF ORIGIN: Fresh-water limestones, gastropods, and algal oöids show that in part the Flagstaff of the Cedar Hills was deposited in a lake, as described in detail by Eardley (1932, p. 405–413). Although the source area of the Flagstaff was essentially the same as for the red North Horn, the Flagstaff is light-colored, indicating the presence of enough organic matter in the lake so that deposition occurred under reducing conditions.

Colton Formation

NOMENCLATURE AND TYPE LOCALITY: The name Colton is applied by Spieker (1936, p. 120–121, 137–139) to the former upper member of the Wasatch formation of central Utah because the name "Wasatch" has been so long and so widely used that its meaning is not always clear. The Wasatch age of the rocks in question has not always been proved. He reviews the past usage and present status of "Wasatch", points out inconsistencies and sources of confusion, and finds it unwise to apply the name to rocks in central Utah until their age is better established. He uses "Wasatch age" as a time division of lower Eocene approximately equal to the Sparnacian of the European time scale. He defines the Colton as "beds in the hills north of Colton between the Flagstaff limestone and the Green River formation" and says it possibly is Wasatch in age. The name Colton is appropriate for the rocks of the same stratigraphic position and similar character in the Cedar Hills because of proximity to the type locality of the Colton and because proof of Wasatch age is no more evident than in the Wasatch Plateau.

DISTRIBUTION: The dominantly red strata overlying the Flagstaff formation in the central part of the Cedar Hills and east of Mount Baldy are identified as the Colton formation, which occurs also as outliers in the northern part of the area.

LITHOLOGY: The Colton consists of fluviatile conglomerates, sandstones, and shales and in general is similar to the upper part of the North Horn formation. It is dominantly red but contains many beds of other colors. Conglomerates up to 10 feet thick contain pebbles, cobbles and boulders of quartzite as much as 10 inches in diameter and pebbles of dark-gray limestone, in a coarse gray or pink matrix of quartz and chert. The sandstones are gray or red and medium- to coarse-grained. The shales generally are red, but some are gray, greenish, or yellowish green.

Red beds on the west side of Big Hollow (sec. 15, T. 13 S., R. 3 E.) are thought to represent the Colton formation, although the exposures are isolated. About nine-tenths of the 247 feet exposed at this locality is variegated shale, and the rest is sandstone and conglomerate.

THICKNESS: The Colton formation is about 580 feet thick in the Water Hollow-Big Hollow divide. Elsewhere in the Cedar Hills the formation is incompletely exposed, either because the base is concealed or the upper part has been eroded.

RELATION TO OLDER FORMATIONS; AGE AND CORRELATION: The Colton rests conformably on and grades upward from the Flagstaff formation, as it does in the near-by Wasatch Plateau. It cannot be traced from the Cedar Hills to the type area because of overlying alluvium or pyroclastic rocks in intervening areas, hence the correlation depends on stratigraphic position and lithology. These strata have not yielded diagnostic fossils, but in the Wasatch Plateau some fresh-water mollusks long considered Wasatch in age have been obtained from them.

Green River Formation

TYPE LOCALITY: Hayden (1869, p. 90; 1873, p. 190) named rocks exposed along Green River at the town of the same name in Wyoming the Green River formation. The formation has been traced discontinuously eastward around the eastern end of the Uinta Mountains into the Uinta Basin of Colorado and eastern Utah.

DISTRIBUTION: An outcrop of the Green River formation 1 to 3 miles wide extends along the eastern margin of the Cedar Hills from Moroni to Whittaker station. Another underlies much of Big Hollow and Uinta Canyon and probably is continuous with the first outcrop at Whittaker, although the connecting area is covered by alluvium. A small outlier of the formation caps the Water Hollow-Big Hollow divide.

LITHOLOGY: The Green River formation consists of light-colored lacustrine shales and limestones with minor sandstones and conglomerates. Although the limestones are predominantly thin-bedded and platy, some of them are massive and gray or drab. The basal bed in the Water Hollow-Big Hollow divide is a white, massive limestone containing small pebbles of quartzite and dark-gray limestone. A massive white limestone ledge at the divide between Big Hollow and Uinta Canyon is partly oölitic.

The shales range from yellowish through red, brown, chocolate brown, gray, greenish gray, and green to dark gray and black. Some of the shale is thin-bedded, but some shows no definite bedding.

The sandstones are dark gray, medium gray, white, light brown, buff, and reddish, fine- to medium-grained, and a few inches to 2 feet thick. Some are friable, concretionary, and micaceous. A bed of sandstone 45 feet thick, which crops out at the base of the low cliffs just southwest of Fairview, is buff and yellow, mostly massive but locally rather thin-bedded, medium- to coarse-grained, and pebbly, friable to well cemented, and in part cross-bedded.

The conglomerates are gray or brown and 5 to 10 feet thick and in general contain smaller pebbles than the conglomerates of older formations. Approximately on the line between secs. 15 and 22, T. 14 S., R. 4 E., the formation contains a conglomerate 5 feet thick, in which are pebbles and cobbles of scoria as much as 1 foot in diameter. The source of these igneous materials is unknown, but the volcanism represented by them may be of Flagstaff age, for beds of dense, white, chalky material found by Spieker (1931, p. 46) in the Flagstaff in the Wasatch Plateau were considered by C. S. Ross as possibly volcanic ash. They suggest that somewhere in central Utah there were lavas of Flagstaff age from which the scoriaceous pebbles and cobbles in the Green River formation could have come.

Because most of the limestones and many of the shales are light-colored or weather white, the cliffs or hogbacks in the lower part of the formation are white. Such cliffs line the west side of San Pitch River from Milburn southwestward past Fairview. Farther southwestward, however, the dip carries these white lower beds below the surface, and the predominant colors are light brown and buff. White cliffs near Spring City, 3 to 8 miles south of the Cedar Hills, may represent a reappearance of the lower strata.

THICKNESS: The base of the Green River formation in the Cedar Hills is exposed only on the Water Hollow-Big Hollow divide, where all but 170 feet of the formation has been eroded.

The thickest measured section, opposite Oak Creek station, totals 350 feet. Because of the southwestward dip, this section and two others measured 5¼ and 6 miles farther southwest in the cliffs along the San Pitch River probably represent different parts of the formation. Their combined thickness is about 700 feet.

A computation based on the width of outcrop indicates a thickness of 4250 feet, but, as the dip is not uniform, this figure may be too high. It is comparable, however, to the 4900 feet reported by Bradley (1931, p. 17) for the Green River in Indian Canyon, in the western part of the Uinta Basin.

RELATION TO OLDER FORMATIONS: The Green River formation lies conformably on the Colton in the Water Hollow-Big Hollow divide, which is the only place in the area where their contact is exposed. The details are obscure, but the contact probably is gradational.

FOSSILS AND AGE: Fossil collections from the Green River formation in the southeastern part of the Cedar Hills include fruits of *Chara*, *Sphaerium* ap., *Unio* cf. *U. washakiensis* Meek *Goniobasis tenera* Hall, *Planorbis utahensis* Meek, cyprid ostracods, and a fish, all identified by Reeside, who dates them as not older than Wasatch nor younger than Eocene.

The Green River formation as recognized in the Western Interior is a mappable unit that transgresses time lines and has age equivalents in formations of contrasting lithology. In Wyoming, Colorado, and northeastern Utah the Green River contains a large flora that Berry

(1924, p. 35) first assigned to an age about equal to middle Uinta but later (1925, p. 360, 361) decided is somewhat older, concluding: "At any rate, the Green River and 'so-called Bridger' floras are well up in the middle Eocene." Subsequently, Brown (1934, p. 45–47) examined both old and new collections of floral material from the formation in western Colorado, eastern Utah, and southwestern Wyoming and found that they most closely resemble the Florissant flora, correlative with the Wind River formation of Wyoming.

According to Spieker (oral communication) and A. A. Baker (personal communication) the Flagstaff, Wasatch, and Green River formations intertongue north and east of Thistle, Utah, and the relations between them are so complicated that stratigraphic position is far from adequate evidence for assigning an age to rocks in the absence of diagnostic fossils. Spieker considers that the rocks of Green River type in this general area may range from late Paleocene to middle Eocene. What part of this time interval is represented by the Green River formation of the Cedar Hills cannot be determined by tracing the beds to other areas because of intervening cover. The solution, therefore, awaits discovery of diagnostic fossils.

Pyroclastic Rocks

DISTRIBUTION: Pyroclastic rocks similar to those widely distributed in other parts of Utah cover a large part of the Cedar Hills. From Morni they extend northeastward to The Cliff, where the outcrop branches. The northeastern branch continues to Whittaker station, and the northwestern one to Salt Creek. These rocks also crop out along Pole Creek, spread out across Spencer and Nebo creeks to Thistle and Bennie creeks, and appear as large outliers at Mount Baldy and at the northern end of the area, and as smaller outliers in Big and Water Hollows.

LITHOLOGY: The following section describes the principal rock types in the pyroclastic sequence:

GENERALIZED SECTION OF PYROCLASTIC ROCKS IN CEDAR HILLS

	Thickness (feet)
(4) Tuff and breccia, yellowish, pinkish, porous, partly vitreous, partly dull (slightly metamorphosed by overlying sill)	50
(3) Tuff, massive, white, with biotite and grains of colorless, milky, and smoky quartz; locally contains dark-red igneous pebbles, rather well-rounded	60
(2) Volcanic conglomerate, massive, coarse, gray, crudely bedded; matrix of gray tuff; pebbles, cobbles, and boulders of dark-colored igneous rock, up to 10 feet across; scoria boulders suggest derivation from surface flows; locally boulders of quartzite and Pennsylvanian limestone are included; light-gray phase occurs above dark gray at mouth of Pole Creek, but reverse relationship holds in lower part of Big Hollow and in Nebo Creek canyon; locally includes gray and green tuff, and red strata nearly free of igneous pebbles.	1100
(1) Green sands, sandstones, and minor conglomerates, partly cemented by calcite but generally friable; well-rounded pebbles up to 2 inches in diameter consist of gray quartzite and igneous rock and are in zones indicating large-scale cross-bedding; green color due to ferrous silicate, probably greenalite	350
Total	1560

Units 1, 3, and 4 have been observed only near The Cliff, but unit 2 is widespread.

On the east side of Salt Hollow (SE¼ sec. 9, T. 11 S., R. 3 E.) red rocks similar to parts of the North Horn and Wasatch formations, but distinguished by the inclusion of scattered igneous pebbles, crop out at about the same elevation as undoubted volcanic conglomerate only a quarter of a mile to the west. They indicate that red rocks were being eroded while the volcanic conglomerate was being deposited, and that locally the red sediments predominated.

Likewise, where Hop Creek leaves the hills (NW¼ sec. 9, T. 13 S., R. 2 E.), gray and brown conglomerate similar to conglomerate in the Indianola appears to be bedded with light-gray volcanic conglomerate. It is thought to belong to the pyroclastics, however, because it is coarser and less well sorted than the Indianola conglomerates and is nearly horizontal, whereas the Indianola at all near-by exposures dips steeply southeastward.

THICKNESS: The thickness of the pyroclastic sequence differs widely from place to place because of irregularities on the unconformable surface below, erosion of the upper part, and differences in the thickess of the several members. The maximum thickness, near The Cliff, is about 1560 feet.

RELATION TO OLDER FORMATIONS: The pyroclastic rocks at different places rest uncon-

formably on all older formations. The contact with the next older formation, the Green River, is generally obscured by debris but is definitely irregular. The pyroclastic rocks clearly overlap all rocks older than the Green River, in places with marked angular discordance.

Like the volcanic rocks of the Oquirrh and Traverse mountains (Gilluly, 1932, p. 42) the pyroclastic rocks of the Cedar Hills were deposited on a surface of considerable relief. Their lower contact is high in the valley wall along the east side of Beaver Dam Creek, but, traced southward, it descends about 1000 feet into Nebo Canyon, rises nearly 1000 feet to the divide between Nebo and Wheatgrass creeks, and then dips down into Wheatgrass Valley. Similar relief was observed near the mouth of the Middle Fork of Pole Creek. The writer believes that, on the whole, this condition is not due to folding but to a rough pre-volcanic topography.

AGE: The pyroclastic rocks of the Cedar Hills have yielded no fossils. They overlie unconformably the Green River formation of Eocene age and are broken by the Basin-Range faults, which are generally considered to be of late Tertiary and Quaternary age. Thus the pyroclastic rocks may be as old as upper Eocene, and as young as Quaternary. Eardley (personal communication) reports the finding of lower Oligocene bones in the Norwood tuff in the north-central Wasatch Mountains. As he has studied the volcanic rocks of both areas and apparently considers them plausible equivalents, an early Oligocene age for the pyroclastic rocks of the Cedar Hills seems possible.

CORRELATION: The pyroclastic rocks of the Cedar Hills are lithologically similar to those of the adjacent Wasatch Plateau (Spieker, oral communication) and are undoubtedly equivalent. Likewise, they are either similar to or continuous with the pyroclastic rocks along the east side of the southern Wasatch Mountains, immediately west of the Cedar Hills, and Eardley (personal communication) considers the latter to be equivalent to his Norwood tuff. Descriptions and photographs also suggest a general similarity between the pyroclastic rocks of the Cedar Hills and the volcanic rocks of such widely separated localities as the High Plateaus to the south (Dutton, 1880), the Oquirrh Mountains to the northwest (Gilluly, 1932, p. 41–43), and the Strawberry Valley area to the northeast (A. A. Baker, oral communication). The volcanic rocks of these areas may be correlatives, but, as most writers believe that volcanic activity occurred in Utah through much of Tertiary time, similarity in origin may suggest unjustifiable correlations.

ORIGIN: The greenish lower unit of the pyroclastic sequence contains rather well-rounded pebbles that suggest deposition by water far from the source of the material, and the irregular bedding suggests a stream environment.

In the volcanic conglomerate, the well-rounded igneous pebbles and subangular to rounded cobbles and boulders, the crude stratification, and poor sorting suggest deposition by large, active streams, much as described by Dutton (1880, p. 78, 224) and Eardley (1933, p. 341). On the other hand, part of the volcanic conglomerate may have originated as a mud flow, as suggested by Eardley (1933, p. 342) for the dark-gray phase in Salt Creek canyon, and by Gilluly (1932, p. 45) for the structureless parts of the volcanic breccia in the Oquirrh Mountains.

The white tuff on top of the Cliff probably is water-laid, as shown by the well-rounded pebbles of dense, compact igneous rock in it. Some of the tuffs at other localities lack such pebbles and therefore may be true accumulations of volcanic ejectamenta. The intrusion of sills caused slight metamorphism in some of the tuffs near the Cliff and produced the pink and yellow varieties of unit 4 of the general section.

Although the very large and abundant igneous boulders suggest that the source of the volcanic materials was near by, the exact site is unknown. The only igneous rock in the entire prevolcanic series of the Cedar Hills is a sill a few inches thick in the Green River formation, and it may have been intruded long after Green River time, perhaps contemporaneously with the materials in the pyroclastic rocks.

Volcanic vents or lava flows have not been discovered in the southern Wasatch Mountains west of the Cedar Hills, on the Gunnison Plateau to the southwest, or on the Wasatch Plateau to the east. Baker (oral communication, 1939) reports several small volcanic vents in the Strawberry Valley region to the north, but it is unlikely that the volcanic debris in the Cedar Hills came from there because the largest

boulders reported by Baker are smaller than some of those in the Cedar Hills.

ORIGINAL EXTENT OF VOLCANIC CONGLOMERATE: The volcanic conglomerate caps The Cliff, Mount Baldy, and several high divides (Pl. 2). Patches too small to map occur on Golden Ridge, near Twin Knolls, and near the head of Black Canyon. At the last two localities the elevation is about 9400 feet. These occurrences of the conglomerate on high parts of the area and its otherwise wide distribution suggest that most of the Cedar Hills was once covered by it.

The volcanic materials probably accumulated first along valleys, but after these had been filled the heavily-laden streams spread debris over the divides, leaving much of the Cedar Hills as a piedmont plain of slight relief.

Eardley (1933, p. 341) has suggested that the present drainage in the region follows about the same lines as the prevolcanic drainage. The fact that the base of the volcanic deposit descends into both Nebo and Pole canyons favors this view. If modern drainage crossed the older drainage, valleys filled by volcanic debris might be exposed in cross section, but none has been discovered.

The re-establishment of the old drainage lines probably means that no important change in the regional slope occurred during the volcanic period. It also implies control of the post-volcanic streams by the old topography, for after cutting through the volcanic debris, which may have been nearly unconsolidated, the streams that encountered the harder rocks in the buried ridges would have tended to migrate laterally in the softer pyroclastic filling of the valleys.

QUATERNARY SYSTEM

Terrace Gravels and Alluvium

Hills of gravel and boulders stand above the present flood plains at several places around the margin of the Cedar Hills. The ridge between Whittaker station and U. S. highway 89 is composed of such material, and a gravel pit has been opened in the north end of it. The ends of spurs east of Fountain Green are either composed of or veneered with such gravels, and about 3 miles north of the town U. S. highway 189 cuts through several small gravel hills rising 10 to 20 feet above the surrounding alluvial flats.

Alluvium floors large areas in Sanpete and Thistle valleys around the margin of the Cedar Hills, and smaller areas in the main valleys within the hills. Much of it was probably dumped by streams in flood and is coarse and poorly sorted. A well record at Fountain Green shows 285 feet of sand, gravel, and clay, all of which is doubtless alluvium (Richardson, 1907, p. 48).

As the terrace gravels are higher they probably are older than the alluvium. In the absence of contrary evidence, both are assigned to the Quaternary.

IGNEOUS ROCKS

Intrusive Rocks

Sills.—A massive sill of dense, dark felsite with few phenocrysts occurs in tuffs near The Cliff and Fountain Green. Its maximum thickness is 25 feet. It overlies pink and yellow tuffs and breccias, which probably owe their coloring to partial metamorphism, and underlies a metamorphosed tuff. A similar sill crops out near the top of The Cliff.

A medium-grained hornblende andesite a few inches thick occurs in the Green River formation in the cliffs along the west side of San Pitch River opposite Oak Creek station. Presumably it is the edge of a sill.

Dikes.—A dull-red felsite dike about 30 feet wide and extending about 50 feet above the road crosses Salt Creek near the mouth of Pole Creek, on the west side of the Cedar Hills (sec. 28, T. 12 S., R. 2 E.). Eardley (1934, Map 11; 1933, p. 343) mapped it and described it as porphyritic andesite similar to the volcanic conglomerate into which it was intruded. If the conglomerate adjacent to the dike was metamorphosed, debris masks the effects of the metamorphism.

Age of intrusive rocks.—The dike on Salt Creek and the sills near The Cliff were intruded into the pyroclastic rocks and therefore are younger. The sill in the Green River formation near Oak Creek station is not older than Green

River. Probably all the intrusive rocks belong to the same period of volcanism.

Eardley (1933, p. 342–344) found several dikes and sills along the west side of the Wasatch Range and concluded that these, the extrusive rocks, and two lamprophyre dikes described by Loughlin (1918) "show petrographic relationships, and all apparently belong to the same eruptive period." The information available suggests a time between early Oligocene and late Tertiary or Quaternary for this period. (*See* Age of Pyroclastic Rocks.)

Metamorphic Rocks

Metamorphosed Tuff

A dense, glassy, red and pink rock thought to be a metamorphosed tuff crops out on high places in the triangle between Fountain Green, Moroni, and The Cliff. There are smaller outcrops west of Big Hollow, west of Spencer Creek (sec. 5, T. 12 S., R. 3 E.), and half a mile south of the forks of Pole Creek. About 3 miles east of Fountain Green the tuff overlies the black sill and ranges from 25 to 45 feet thick within half a mile.

This rock contains abundant flakes of biotite, some quartz, and a colorless mineral which may be feldspar (sanidine?). Some specimens contain many white spots of soft material, possibly kaolinized feldspar. A few rounded bodies of quartzite are interpreted as metamorphosed quartzite pebbles. Any igneous pebbles in the tuff before intrusion have been altered beyond recognition. Small black stringers occur locally in the red rock and may represent streams of the intruding material.

The rock resembles rhyolite but is thought to be a metamorphosed tuff because intermediate rock types suggest gradation from white and gray tuff containing biotite and crystals of quartz to the "rhyolitic" rock, which also contains biotite and quartz. The intermediate types occur below a dark sill, consist of pink and yellow semivitrified tuff and breccia, and grade down into gray-white tuff. The "rhyolite" is above the sill and is more dense and vitreous, doubtless because the metamorphic effect of the sill was more intense upward.

Structural Geology

General Structural Features

The Cedar Hills are in the boundary zone between the Colorado Plateaus and the Great Basin. To the east is the Wasatch Plateau, an

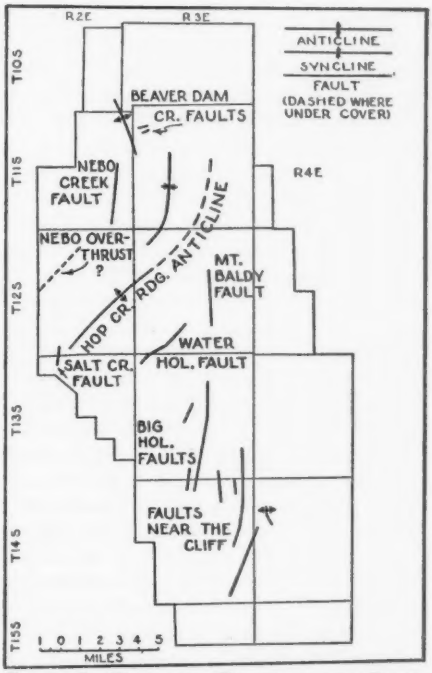

Figure 3.—Folds and Faults of the Cedar Hills, Utah

elevated block of nearly horizontal Mesozoic and Tertiary sediments bounded on the east by an erosional cliff and on the west by the Wasatch monocline. To the west and north are the southern Wasatch Mountains, an anticlinal range consisting chiefly of folded Paleozoic sediments. South of the Wasatch Range is the Gunnison Plateau, composed of Mesozoic and Tertiary rocks folded in a broad syncline whose eastern limb is sharply overturned (Spieker, personal communication).

In most parts of the Cedar Hills the rocks dip 25° S.E., or less, but in the west-central part

dips of 75° to 90° SE. prevail. The rocks of the western, central, and northern parts of the area have been folded, and those of the southern part have been broken by normal faulting (Fig. 3).

Folds

Southern Wasatch Mountains anticline.—Some of the steeply dipping rocks in the western portion of the Cedar Hills belong to the anticline of the southern Wasatch Mountains. As described by Eardley (1934, p. 379–381), the southern 8 or 10 miles of this anticline is "practically isoclinal and overturned" and is overthrust eastward. From the trace of the thrust plane along the flank of the range eastward into the Cedar Hills as far as Hop Creek, there is a normal stratigraphic succession from Arapien (Jurassic) to Indianola (Cretaceous); also, the general strike is northeastward throughout, and the dip diminishes from 90° or overturned at the foot of the range to 60° SE. These facts suggest that the Indianola of the western Cedar Hills is part of the eastern limb of the anticline in the block under the thrust. Not all of this block, however, is genetically related to the thrust, for the strong angular unconformity between the Indianola and the steeply dipping Price River at the northeastern end of Hop Creek Ridge shows that the present attitude of beds there is due to two separate movements. Hence, only the part farther west was folded with the Wasatch Mountains anticline.

Hop Creek Ridge anticline.—The Hop Creek Ridge anticline, in the west-central part of the Cedar Hills, strikes northeastward approximately along Hop Creek Ridge, for which it is named. It affects the Flagstaff and older formations, is partly asymmetrical with the steeper limb on the southeast side, and is about 11 miles long. Along the northeastern 5 miles, only the northwestern limb is exposed; along the southwestern 6 miles, the southeastern limb is well exposed and the northwestern limb poorly. The general structure is indicated by the following dips measured roughly along the line of cross section C-D (Pl. 2), beginning at the northwest: west of Left Fork of Pole Creek (Price River formation), 10° NW.; Middle Fork of Pole Creek (lower North Horn formation), 28° NW.; head of Right Fork of Pole Creek (lower North Horn), about horizontal—axis of fold; mouth of Rough Canyon (Price River and lower North Horn), 75° SE.; northwestern Water Hollow (upper North Horn and Flagstaff), 10°-15° SE.

The post-Indianola erosion surface, which truncates Indianola strata at the northeastern end of Hop Creek Ridge and may once have been reasonably flat, is arched over the ridge. It crops out along the Middle Fork of Pole Creek at an elevation of about 7700 feet, rises to 8000–8500 feet at the head of the Right Fork of Pole Creek, and descends to 6700–8000 feet along the southeastern base of Hop Creek Ridge. Along the crest of Hop Creek Ridge this surface has been destroyed by removal of the Price River, but it probably was once higher than the ridge, which now ranges from 8400 to 8700 feet above sea level. Thus the arching of the erosion surface is between 1000 and 2000 feet.

The northwestern limb of the anticline steepens northeastward from a dip of 28° NW. along the Middle Fork of Pole Creek to 57° NW. along Wheatgrass Creek (Pl. 2, cross section A-B), where the opposing limb is covered by pyroclastic rocks and there is no way of knowing whether or not this part of the fold is symmetrical.

Folds in Nebo Creek basin.—In the Nebo Creek basin the Hop Creek Ridge anticline is flanked on the northwest by a shallow syncline and a second and smaller anticline. The common limb of the syncline and of the Hop Creek Ridge anticline is revealed by the North Horn formation, which dips 40° to 57° NW. along Wheatgrass Creek. Within half a mile to the northwest the dip decreases to 25° or 30° NW. These dips are opposed by northeast dips of 16° to 25° in Golden Ridge, about 2 miles to the west, and also on the east side of Beaver Dam Creek, about 4½ miles to the northwest. The axis of the syncline probably lies in the valley of Pages Fork, and, as the opposing dips would intersect roughly at right angles, it probably plunges northward, going under pyroclastic rocks north of Nebo Creek.

The second anticline lies along the ridge west of Beaver Dam Creek and is about 3 miles long. A dip of 28° W. in the ridge is opposed by

dips of 21° and 25° NE. on the east side of the creek. This fold was not observed in the rocks south of Nebo Creek.

Other folds.—The Flagstaff and younger formations in Water and Big hollows dip southeastward at angles up to 10°, with slight warpings but without folds comparable to those described above. An open anticline in the Green River formation about 1½ miles southeast of The Cliff (sec. 7, T. 14 S., R. 4 E.) trends northward and probably disappears under the pyroclastic rocks. Dips on the flanks of it range from 15° SE. to 22° W.

Faults

General statement.—Most of the faults observed in the Cedar Hills affect the nearly horizontal or gently to moderately tilted rocks. They are partly exposed and partly inferred from stratigraphic and topographic evidence. They strike roughly north-south, are relatively straight, and probably dip at high angles.

Only one possible fault was found in the steeply tilted rocks of the Pole Creek-Hop Creek area. If there are others, they are bedding-plane faults or have brought similar lithologic types against each other and therefore have escaped detection.

The faults are described in the order of the age of the rocks involved.

Nebo overthrust.—The Nebo overthrust may be present under cover in the western part of the Cedar Hills. Eardley (1934, Map 11, p. 381–383) traced the thrust plane around the southern end of Mount Nebo into Salt Creek valley as far as the mouth of Cow Hollow, where it passes under the Price River formation. He suggested that the fault dies out to the northeast, in the area of this report.

Salt Creek fault.—A possible fault about half a mile east of the lower end of Salt Creek (NW¼ sec. 4, T. 13 S., R. 2 E.) is suggested by erratic strikes and dips. Within a quarter of a mile, the strike and dip change from N. 48° E., 26° SE., to N-S., 70° E., and then to N. 32° W., 40° NE. The strike of this fault is estimated as N. 5–10° E. Only the Indianola group is involved, hence the age of the fault is known only to be post-Indianola.

Nebo Creek fault.—A north-south fault whose effects are seen in the Price River formation crosses Nebo Creek and Black Canyon about a quarter of a mile west of their junction (sec. 24, T. 11 S., R. 2 E.) and is exposed in a spur about 1½ miles to the south. Near the northern end, it is downthrown about 160 feet on the east side. East of the fault, the Price River formation dips about 4½° W., but on the other side it dips 20° W.

Beaver Dam Creek faults.—A small graben about a quarter of a mile wide is bounded by faults that dislocate thick ledges of the North Horn and Flagstaff formations on the east side of Beaver Dam Creek. The southern fault is downthrown about 60 feet on its north side, and the other is downthrown about 200 feet on its south side. The faults were seen only close to the creek but are thought to strike roughly east-west. They apparently did not affect the pyroclastic rocks capping the ridge east of the creek.

Water Hollow fault.—A fault about 3 miles long strikes northeastward along ravines in the northwestern part of Water Hollow. Near its northeastern end, widely divergent dips in closely spaced exposures of platy limestone indicate shattering of the Flagstaff formation. Faulting is also suggested by several small springs that issue along the otherwise dry ravines.

Physiographic evidence suggests that the upthrow is on the southeast side. Stream 1 (Fig. 4) heads in a broad, open saddle near the crest of the ridge that parallels the fault on the southeast. The gullies at 2 are aligned as if they were its headwaters, but they join ravine 3 and drain southwestward into stream 4. These conditions suggest that stream 1 has been beheaded by the faulting, in which case the upthrow would be on the southeast.

Mount Baldy fault.—A fault or zone of faults downthrown on the east appears to be partly responsible for the relative height and prominence of Mount Baldy. The eastern face of the mountain is an escarpment that probably originated as a fault scarp but has been modified and shifted westward by glacial carving or by gradual sapping by meltwater from snow banks, which sometimes last until late July. Pyroclastic rocks are exposed in the escarpment, and in the basin to the east these rocks alternate

with conglomerates and limestones of North Horn or Colton type, thus suggesting a zone of faults. The total displacement probably is several hundred feet.

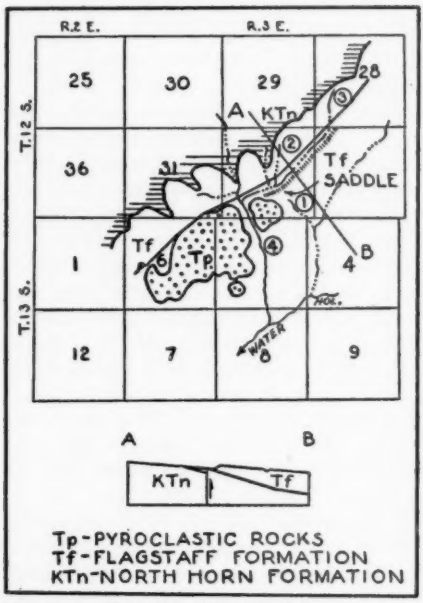

FIGURE 4.—DRAINAGE CHANGES CAUSED BY THE WATER HOLLOW FAULT

Faults near The Cliff.—The main fault at The Cliff is about 5 miles long and is downthrown on the east. The peak itself is capped in part by the igneous sill and the metamorphosed tuff, and these rocks also occur about 500 feet lower on the next ridge to the east. The precipitous east face of The Cliff descends into a narrow, north-south trough that probably owes its existence to more rapid erosion in the shattered zone. Platy limestones in the northern part of the trough indicate the faulting by wide divergence of strikes and dips within short distances.

A branch fault cuts across the top of The Cliff, bringing the white tuff in contact with an igneous sill. It is downthrown between 50 and 130 feet on the east, and the fault plane dips about 58° E.

About 1 mile west of The Cliff the southward-facing wall of a small canyon exposes the massive tuff of the pyroclastic sequence faulted down against the greenish basal member. The volcanic conglomerate, which occurs between these two at some near-by localities, appears to be absent, hence the throw may be about equal to the thickness of the conglomerate—possibly 1000 feet The downthrow is on the east, and the strike probably is northward.

The Spring Hollow that heads about 3 miles south of The Cliff probably follows the shattered zone of a normal fault. It trends approximately in the same direction as faults near The Cliff, and along it the platy limestones of the Green River formation are highly disturbed.

Big Hollow faults.—A scissors fault about 5½ miles long follows the western branch of Big Hollow. Near the southern end (approximately sec. 34, T. 13 S., R. 3 E.) the rocks west of the fault belong to the upper part of the pyroclastics. They are opposed on the east by the basal greenish sand of the pryoclastics, hence the downthrow is on the west. Near the northern end, however, the strata west of the fault are the Colton formation, opposed on the east by beds of the Green River, and the downthrow is on the east. Along the middle portion of the fault, the direction of throw is not apparent, but the fault is indicated by the relative straightness of the east side of the flat-bottomed valley in contrast with the irregularity of the west side, whose spurs and re-entrants suggest a mature topography partly buried in alluvium.

A fault about 1 mile long in sec. 33, T. 13 S., R. 3 E., and adjacent sec. 4 displaces pyroclastic strata, a sill, and metamorphosed tuff. It strikes approximately north and is downthrown on the east; the throw near the southern end is about 150 feet, decreasing northward to about 60 feet.

A short fault downthrown on the east side has been mapped along a steep scarp in secs. 16 and 21, T. 13 S., R. 3 E. A patch of pyroclastic rock crops out on the east and lower side, and the Colton formation on the west. The amount of throw is unknown.

Epochs of Folding and Faulting

General considerations.—Recent investigations have shown that the central North American Cordillera was involved in at least 9 and perhaps as many as 15 compressional move-

ments in Mesozoic and Cenozoic time; that folding advanced from west to east, reaching central Utah in mid-Cretaceous time; and that the effects of the orogenies were local (Spieker, 1939; Eardley, 1941). Of the compressional movements that affected central Utah between Upper Jurassic (?) and late Tertiary time, four are recognized in the Cedar Hills, and in addition the area was involved in the normal faulting of the Basin-Range disturbance.

Pre-Indianola orogeny.—The first orogeny affecting the Cedar Hills is inferred from the coarse conglomerates of the Indianola, which indicate large, active streams with high gradients and suggest that erosion had been accelerated by the folding or uplifting of mountains. That the orogeny occurred west of the Cedar Hills is shown by the eastward gradation of the Indianola into the marine Mancos shale (Spieker, 1946, p. 127–128). As no angular discordance is apparent between the Indianola and the underlying Arapien where they are rather well exposed on the west side of the Gunnison Plateau, most of the movement probably occurred outside the Cedar Hills-Gunnison Plateau area.

This orogeny occurred after deposition of the Arapien shale (Upper Jurassic) and before the Indianola (Upper Cretaceous), hence at the youngest may be about middle Cretaceous and equal to Spieker's early Colorado orogeny. On the other hand, if the lower Indianola includes equivalents of the Morrison formation (Upper Jurassic), as suggested in the section on Fossils and age of the Indianola, the "pre-Indianola orogeny" would necessarily be of Upper Jurassic age, probably correlative with the early Uinta (early Morrison?) uplift of Eardley (1941, p. 1899; personal communication).

Likewise, the possibility that the lower part of the Indianola is Lower Cretaceous, as already discussed, means that the orogeny may be of early Cretaceous age, but the general absence of Lower Cretaceous rocks in central Utah makes this interpretation unlikely. From these considerations it is apparent that more accurate dating of unit 1 of the Indianola is needed before the age of the orogeny can be determined closely.

Late Indianola orogeny.—The pre-Indianola orogeny was followed by a period of crustal quiet during which the marine sandstone member (unit 2) of the Indianola was deposited. Then renewed folding or uplift caused the streams to bring the great quantities of coarse debris in the upper conglomeratic member of the Indianola (unit 3). This late Indianola orogeny is the second one recorded in the Cedar Hills. It occurred during Colorado time and correlates with the late Colorado (Niobrara) orogeny of Spieker (1939).

Montana (early Laramide) orogeny.—The folding of the southern Wasatch Mountains and the overthrusting of Mount Nebo, in which the western part of the Cedar Hills was involved, probably occurred during Montana time. Clearly this orogeny is not older than Colorado, for it affected the Indianola group, which contains fossils of Colorado age. Furthermore, the fossiliferous strata of the Indianola are overlain by more than 2000 feet of Indianola beds that may even be of early Montana age, and these also were folded.

If the Price River formation is correctly identified in the Cedar Hills, this orogeny is not younger than Montana, for the Nebo overthrust disappears under the Price River, and the truncated limb of a fold contemporaneous with it is overlain unconformably by the Price River at the northeastern end of Hop Creek Ridge (discussed in section on Relation of Price River to older formations). This orogeny is equivalent to the late Montana (early Laramide) orogeny of Spieker (1939) and probably, also, to the first major uplift of the Uinta Mountains (Eardley, 1941).

The structures produced in this orogeny were a large anticline occupying the site of the southern Wasatch Mountains, and a smaller one extending southward from the central Cedar Hills into Sanpete Valley as suggested by Spieker (oral communication) on the basis of observations on either side of the valley in the Wasatch and Gunnison plateaus. The smaller anticline may have branched from the larger one at the northeastern end of Hop Creek Ridge where Indianola beds are truncated by the Price River formation. The truncated fold exposed there may be interpreted as the remains of the northeastern end of the syncline between the two anticlines (Fig. 5 A) or as part of an east-

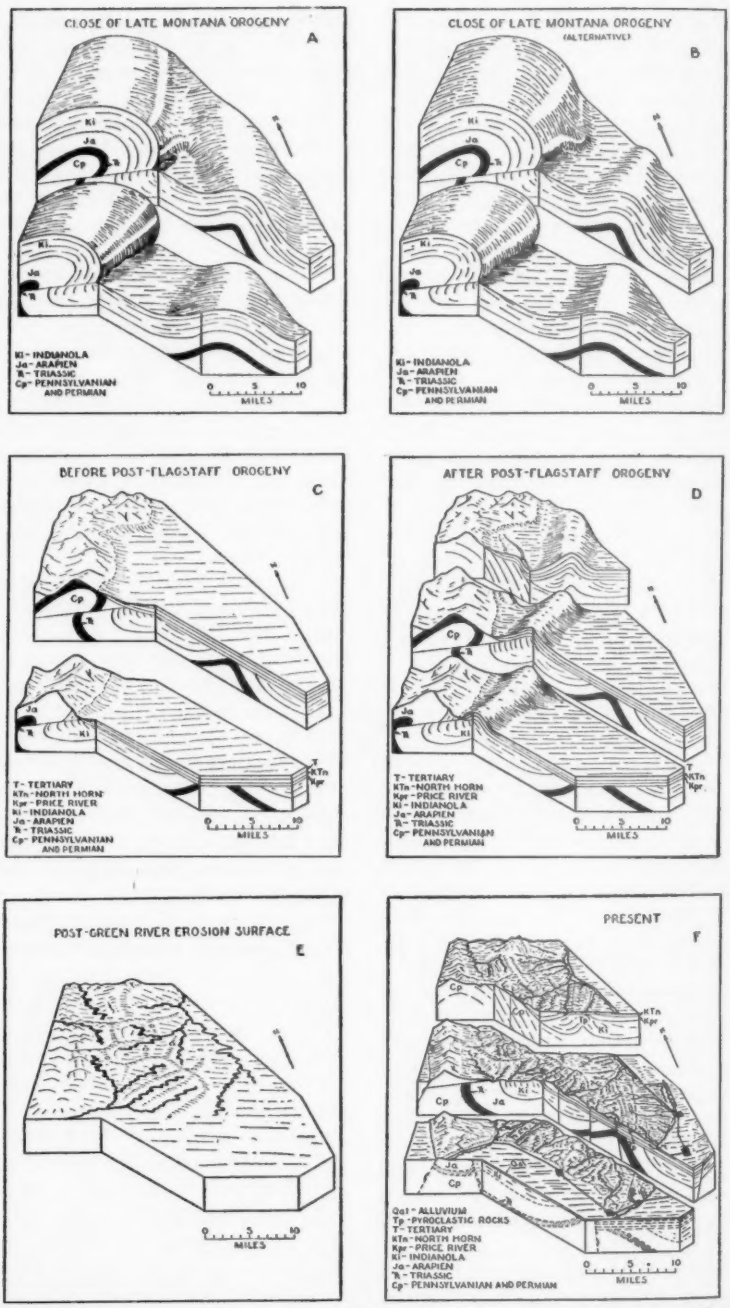

FIGURE 5.—STRUCTURAL AND PHYSIOGRAPHIC DEVELOPMENT OF THE CEDAR HILLS AND VICINITY

ward bulge or wrinkle on the flank of the Wasatch Mountains anticline (Fig. 5 B).

Post-Flagstaff disturbance.—The disturbance that produced the Hop Creek Ridge anticline and associated structures in the Nebo Creek basin (Fig. 5 C, D) cannot be dated so readily as the Montana orogeny. It tilted the Price River and North Horn formations of the Hop Creek basin to steep angles and may have produced the dips of 10° to 15° SE. in the Flagstaff formation in the northwestern part of Water Hollow. Therefore, this disturbance probably is post-Flagstaff and it may have been responsible for the influx of the clastic sediments of the Colton formation.

On the other hand, the Colton and Green River formations may have been involved in this disturbance, for their gentle dips of about 5° SE. across Big Hollow could have been caused either by it or by the Basin-Range disturbance. Also, the small anticline in the Green River formation southeast of The Cliff (sec. 7, T. 14 S., R. 4 E.), which is post-Green River and prevolcanic in age, may be due to this disturbance.

As the Hop Creek Ridge anticline is overlapped by pyroclastic rocks, the disturbance must have occurred between the close of Flagstaff time and the volcanism. Thus it may be either early Eocene or—if the pyroclastic rocks are of early Oligocene age—late Eocene or very early Oligocene. It probably is equivalent to the orogeny which Spieker (personal communication) says occurred at some time after the middle Eocene and before the later volcanics, probably in mid-Tertiary time; and perhaps to the upper Eocene gentle folding that Eardley (1941) reports for the north-central Wasatch Mountains.

Basin-Range disturbance.—The normal faults in the Cedar Hills displace rocks ranging from the Indianola group (Cretaceous) to the pyroclastic beds, but all may have been formed in the same disturbance, probably the Basin-Range disturbance.

This disturbance cannot be dated closely for the Cedar Hills because the age of the youngest rocks involved—the pyroclastics—is not accurately known. If they are early Oligocene, the disturbance could have occurred at any subsequent time. According to the usual interpretation, normal faulting was widespread and of long duration in Utah and adjacent areas during late Tertiary time. King (1933, p. 40) reports pre-Miocene faulting in the Great Basin in Nevada, and others (Dutton, 1880, p. 35; Gilbert, 1928, p. 67; Gilluly, 1932, p. 86; Nolan, 1935, p. 64) have suggested normal faulting in the region at intervals from Miocene or Pliocene until very recent time. Nolan (1935, p. 55, 64) has also recognized four stages of normal faulting that range from early Eocene, or earlier, to late Eocene or early Oligocene. Whether the normal faulting in the Cedar Hills corresponds to one of these, or is equivalent to his late normal faulting, which he suggests is upper Pliocene, remains an open question.

Physiography

The Cedar Hills exhibit a variety of peaks, ridges, canyons, open valleys, terraces, and drainage peculiarities suggesting a significant physiographic history. One of the first recorded events was the development of a late-mature erosion surface (Fig. 5 E) that cuts across strata ranging from the Indianola group to the Green River formation but goes under the pyroclastic rocks. This surface is represented by the Nebo Creek divide and subsidiary ridges in the Nebo Creek, Pole Creek, Hop Creek, and Water Hollow areas. It probably was highest near the head of Salt Creek and, sloping eastward, was diversified by hills up to 100 feet high and by escarpments held up by the more resistant rocks. Mount Nebo and Mount Santaquin, of the Wasatch Range, probably rose about 2500 feet above it, and Twin Knolls were about 500 feet above it. If the pyroclastic rocks are early Oligocene, this erosion surface was developed in late Eocene time. When it had reached late maturity, a new cycle of erosion was begun, streams dissected it to produce an early mature topography, and then the pyroclastic rocks were spread over the area. The epoch of pyroclastic sedimentation left the area a piedmont plain of slight or moderate relief. The streams again attacked the area and carved the present rugged topography (Fig. 5 F).

Conclusion

The Cedar Hills of central Utah, in the boundary zone between the Colorado Plateaus and the Great Basin, contain a section of Upper

Cretaceous and lower Tertiary sediments estimated to range from 21,000 to 28,000 feet in thickness, capped by a sequence of Tertiary pyroclastic rocks with a maximum thickness of about 1500 feet. The Cretaceous and Tertiary rocks are mostly continental but include a zone of fossiliferous marine sandstones and associated conglomerates that date the Indianola group as Upper Cretaceous (Colorado). These fossiliferous beds also make it possible to date with some accuracy the orogenic movements that affected the area.

The orogenies are recorded by the thick series of coarse conglomerates indicating rejuvenation of streams probably traceable to folding or uplift; by an angular unconformity; and by observed folds and faults. The first recorded orogeny is thought to have occurred in mid-Cretaceous time, possibly early Colorado. The second is of late Colorado (Niobrara?) age, and the third Montana. The fourth cannot be dated closely and may be either early Tertiary (Wasatch) or mid-Tertiary. The fifth and latest is the Basin-Range disturbance of late Tertiary or Quaternary age.

The recorded physiographic history appears to include parts of at least three cycles of erosion, the first of which may be approximately mid-Tertiary, and one burial beneath pyroclastic rocks.

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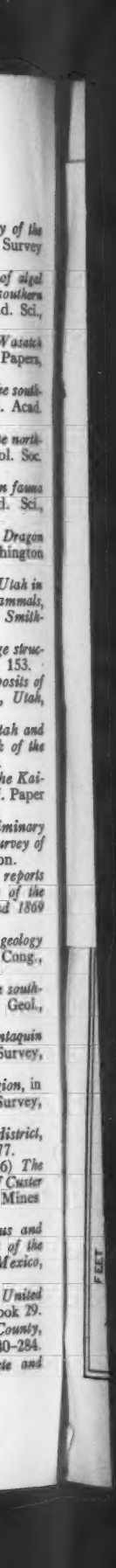

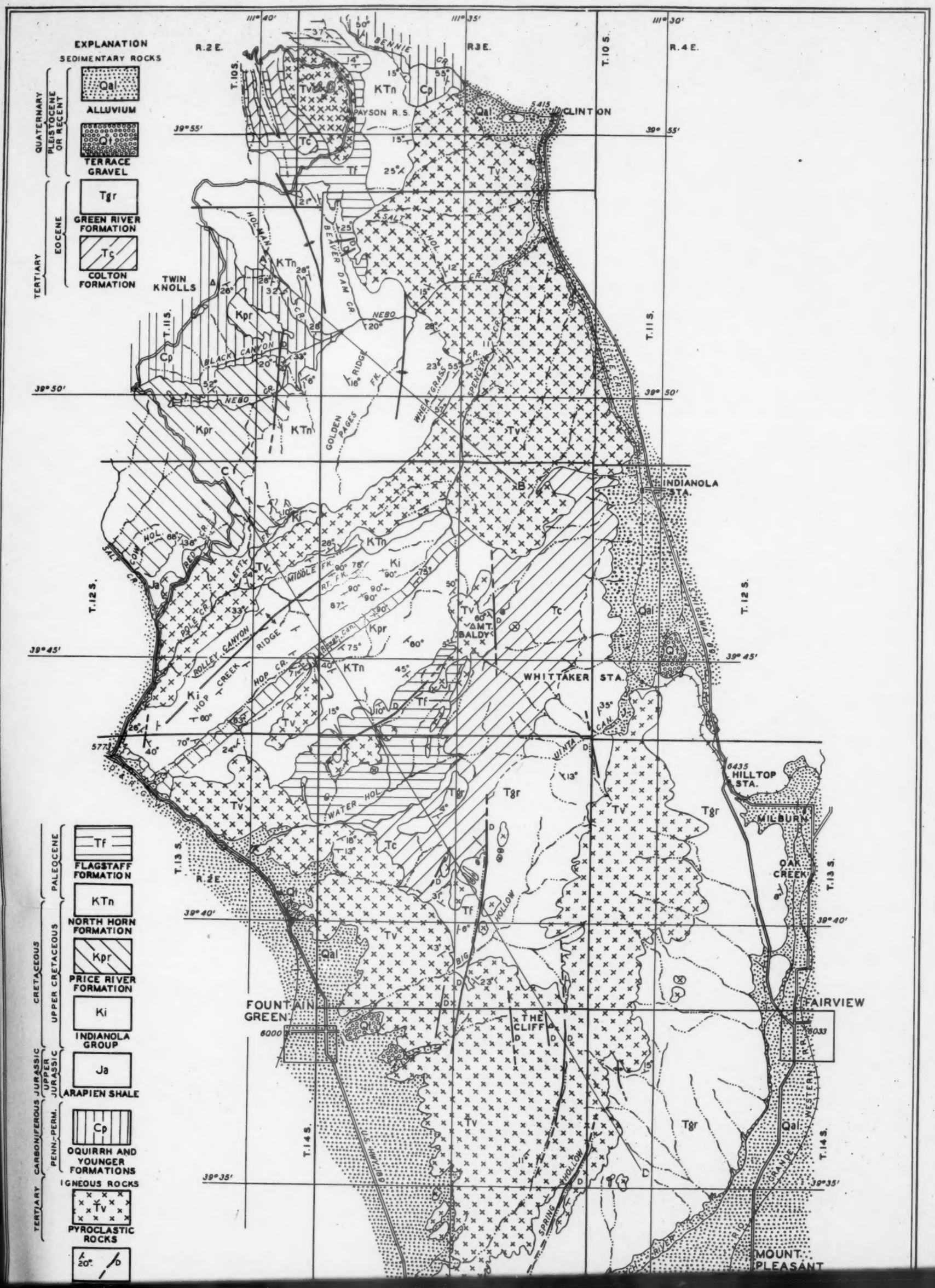
EXPLANATION
SEDIMENTARY ROCKS
QUATERNARY
PLEISTOCENE OR RECENT
Qal
ALLUVIUM
Qt
TERRACE GRAVEL
TERTIARY
EOCENE
Tgr
GREEN RIVER FORMATION
Tc
COLTON FORMATION
PALEOCENE
Tf
FLAGSTAFF FORMATION
KTn
NORTH HORN FORMATION
CRETACEOUS
UPPER CRETACEOUS
Kpr
PRICE RIVER FORMATION
Ki
INDIANOLA GROUP
UPPER JURASSIC
Ja
ARAPIEN SHALE
CARBONIFEROUS JURASSIC
PENN.-PERM.
Cp
OQUIRRH AND YOUNGER FORMATIONS
IGNEOUS ROCKS
Tv
PYROCLASTIC ROCKS
111° 40'
111° 35'
111° 30'
R.2 E.
R.3 E.
R.4 E.
T.10 S.
T.11 S.
T.12 S.
T.13 S.
T.14 S.
39° 55'
39° 50'
39° 45'
39° 40'
39° 35'
CLINTON
PAYSON R.S.
TWIN KNOLLS
INDIANOLA STA.
WHITTAKER STA.
HILLTOP STA.
MILBURN
OAK CREEK
FAIRVIEW
FOUNTAIN GREEN
THE CLIFF
MOUNT. PLEASANT
MT. BALDY
BENNIE
NEBO
BLACK CANYON
GOLDEN PAGES
WHEAT GRASS
SPENCER
BEAVER DAM CR.
ROLLEY CANYON
HOP CREEK
RIDGE
HOLLOW
SPRING
WATER HOL.
U.S. HIGHWAY 89
D. & R. G. W. R.R.
UTAH RAILWAY
GRAND RIVER
5415
6435
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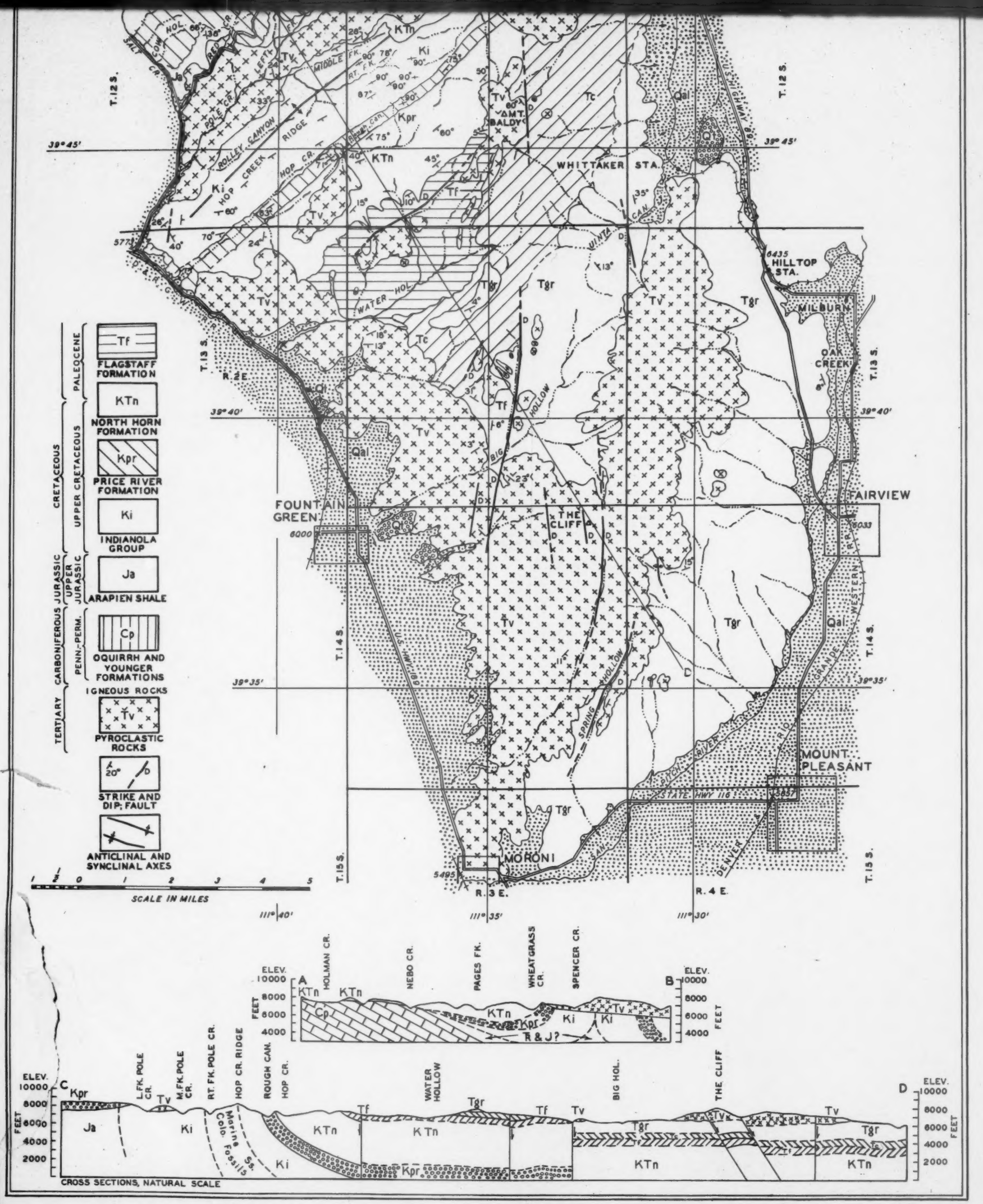
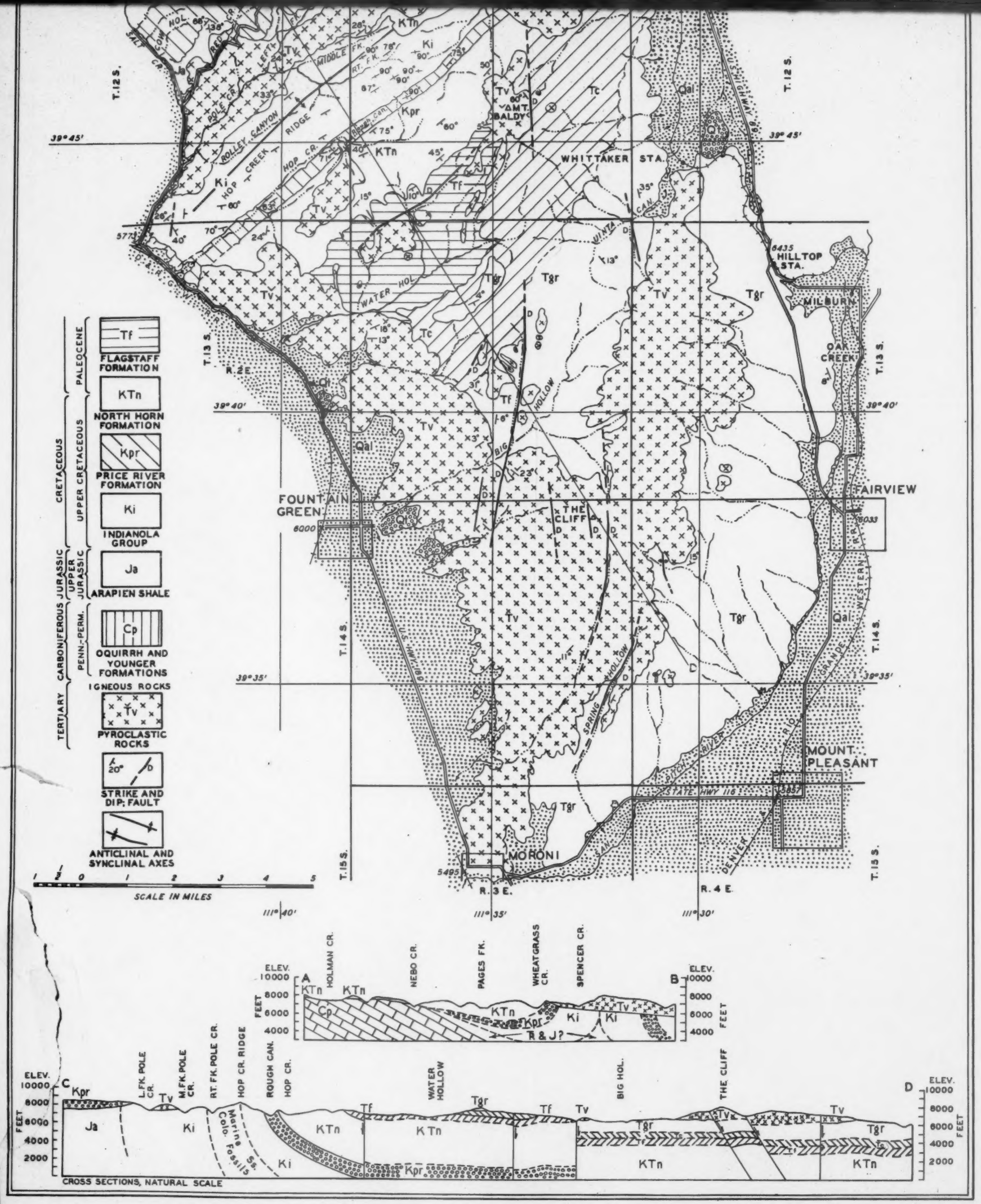


GEOLOGIC MAP OF THE CEDAR HILLS, UTAH



GEOLOGIC MAP OF THE CEDAR HILLS, UTAH

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NORMAN, OKLAHOMA.

RECEIVED BY THE SECRETARY OF THE SOCIETY, MAY 24, 1949

LEGEND

METAMORPHIC AND IGNEOUS ROCKS

DEVONIAN

Late?

Areas in which felsic dikes are abundant
(In Vermont the dikes are medium-grained to fine-grained Ryegate granodiorite, composed of oligoclase, quarts, biotite, microcline, and muscovite. Around the French Pond granite the dikes are pink biotite granite composed of potash feldspar, oligoclase, quarts, and biotite.)

Ryegate granodiorite
(Medium-grained to coarse-grained, light-gray quartz monzonite, granodiorite, and quartz diorite composed of oligoclase, quarts, microcline, biotite, and muscovite; porphyritic granodiorite, prg, near Blue Mountain.)

Haverhill granodiorite
(Medium-grained white or gray granodiorite, locally weakly foliated, composed of oligoclase-andesine, quarts, microcline, biotite, and muscovite.)

French Pond granite
(Heterogeneous body of granite. The most extensive phase, fpgc, is a porphyritic to coarse-grained, pink biotite, granite composed of potash feldspar, oligoclase, quarts, and biotite. A less common phase, fpg, is a pink to gray, medium-grained to fine-grained biotite granite composed of potash feldspar, quarts, oligoclase or albite, biotite, and muscovite.)

bg

Bethlehem gneiss
(Gray foliated granular quarts monzonite or granodiorite composed of quarts, oligoclase-andesine, potash feldspar, biotite, and muscovite.)

Areas in which metamorphosed mafic dikes are abundant
(Pattern indicates areas where mafic dikes are abundant, but they are present throughout the quadrangle. East of the Ammonoosuc thrust the mafic dikes are amphibolite. West of the Ammonoosuc thrust near Haverhill they are metadiabase and green schist. Elsewhere between the Ammonoosuc thrust and the biotite isograd they are green schist. Around the biotite isograd they are albite-epidote amphibolite. West of the garnet isograd they are amphibolite.)

lhdc

Leighton Hill dike complex
(Numerous dikes of green schist and albite-epidote amphibolite constitute the only outcrops.)

Lower

Ol

Littleton formation
(Mica schist, mica-quarts schist, and micaceous quartzite; biotite and garnet porphyroblasts locally common although in many instances they have been chloritized.)

The lower part of the Littleton formation, all of the Clough formation, all of the Fitch formation, and the upper part of the Partridge formation are missing in this quadrangle because of the Northey Hill thrust.

Op

Partridge formation

METAMORPHIC ZONES

Metamorphic zones shown by isograds, hachured on high-intensity side. Rocks on east side of Ammonoosuc thrust are in staurolite zone, rocks on west side are in chlorite zone.

CONTACTS

Accurate

Approximate and diagrammatic due to poor exposures

Indefinite as sharp contact is lacking

SPECIAL SYMBOLS

(Strike and dip symbols represent only a small percentage of the field observations)

Strike and dip of bedding

Strike of vertical bedding

Strike and dip of schistosity

Strike of vertical schistosity

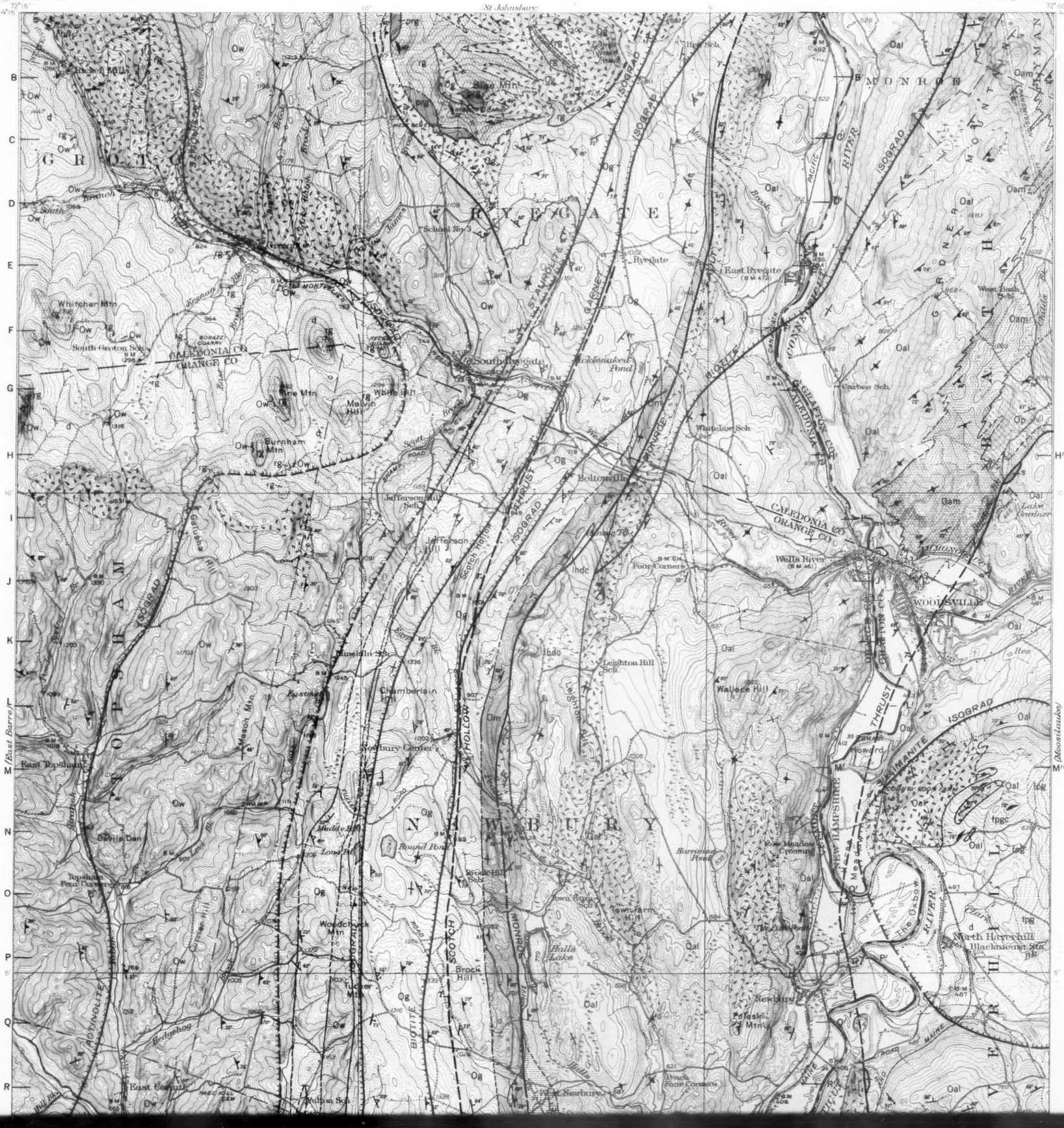
Strike and dip of slip cleavage

Strike and dip of schlieren or foliation in plutonic rocks

Fault

T denotes overthrust side of thrust faults where known.

Silicified fault zone



LEGEND

METAMORPHIC AND IGNEOUS ROCKS

Areas in which felsic dikes are abundant
(In Vermont the dikes are medium-grained to fine-grained Ryegate granodiorite, composed of oligoclase, quartz, biotite, microcline, and muscovite. Around the French Pond granite the dikes are pink biotite granite composed of potash feldspar, oligoclase, quartz, and biotite.)

rg | prg

Ryegate granodiorite
(Medium-grained to coarse-grained, light-gray quartz monzonite, granodiorite, and quartz diorite composed of oligoclase, quartz, microcline, biotite, and muscovite; porphyritic granodiorite, prg, near Blue Mountain.)

hg

Haverhill granodiorite
(Medium-grained white or gray granodiorite, locally weakly foliated, composed of oligoclase-andesine, quartz, microcline, biotite, and muscovite.)

fpg | fpgc

French Pond granite
(Heterogeneous body of granite. The most extensive phase, fpgc, is a porphyritic to coarse-grained, pink biotite, granite composed of potash feldspar, oligoclase, quartz, and biotite. A less common phase, fpg, is a pink to gray, medium-grained to fine-grained biotite granite composed of potash feldspar, quartz, oligoclase or albite, biotite, and muscovite.)

bg

Bethlehem gneiss
(Gray foliated granular quartz monzonite or granodiorite composed of quartz, oligoclase-andesine, potash feldspar, biotite, and muscovite.)

Areas in which metamorphosed mafic dikes are abundant
(Pattern indicates areas where mafic dikes are abundant, but they are present throughout the quadrangle. East of the Ammonoosuc thrust the mafic dikes are amphibolite. West of the Ammonoosuc thrust near Haverhill they are metadiabase and green schist. Elsewhere between the Ammonoosuc thrust and the biotite isograd they are green schist. Around the biotite isograd they are albite-epidote amphibolite. West of the garnet isograd they are amphibolite.)

lhdc

Leighton Hill dike complex
(Numerous dikes of green schist and albite-epidote amphibolite constitute the only outcrops.)

DEVONIAN — *Late?*

DEVONIAN — *Lower*

Dl

Littleton formation
(Mica schist, mica-quartz schist, and micaceous quartzite; biotite and garnet porphyroblasts locally common although in many instances they have been chloritized.)

The lower part of the Littleton formation, all of the Clough formation, all of the Fitch formation, and the upper part of the Partridge formation are missing in this quadrangle because of the Northey Hill thrust.

Op

Partridge formation

METAMORPHIC ZONES

Metamorphic zones shown by isograds, hachured on high-intensity side. Rocks on east side of Ammonoosuc thrust are in staurolite zone, rocks on west side are in chlorite zone.

CONTACTS

Accurate

Approximate and diagrammatic due to poor exposures

*Indefinite as sharp contact is lacking

SPECIAL SYMBOLS

(Strike and dip symbols represent only a small percentage of the field observations)

Strike and dip of bedding

Strike of vertical bedding

Strike and dip of schistosity

Strike of vertical schistosity

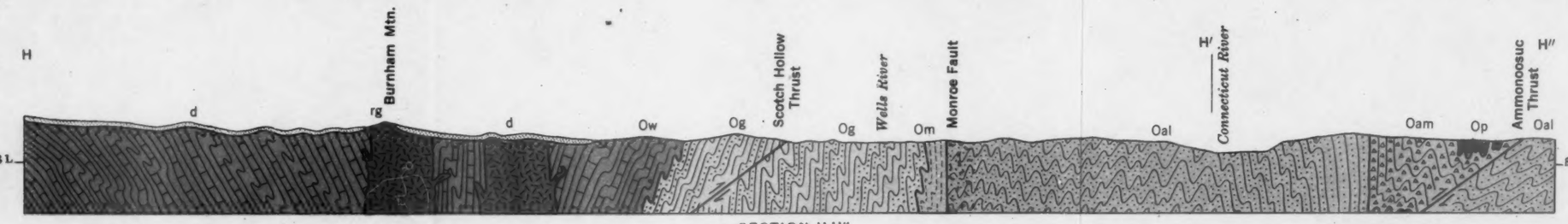
Strike and dip of slip cleavage

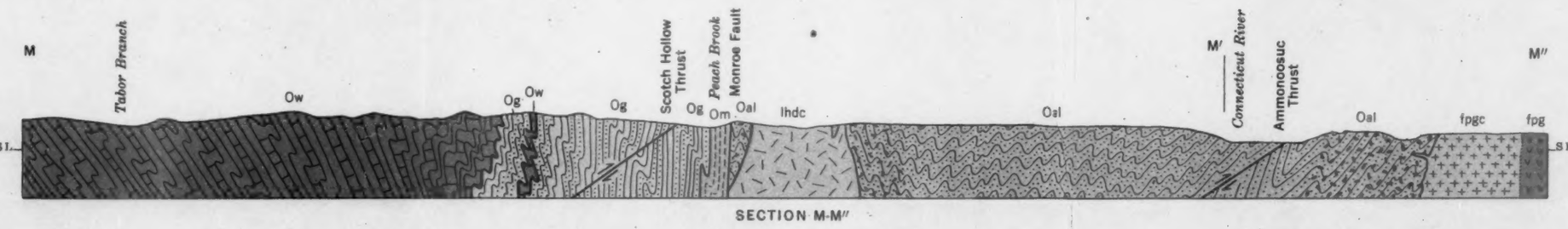
Strike and dip of schlieren or foliation in plutonic rocks

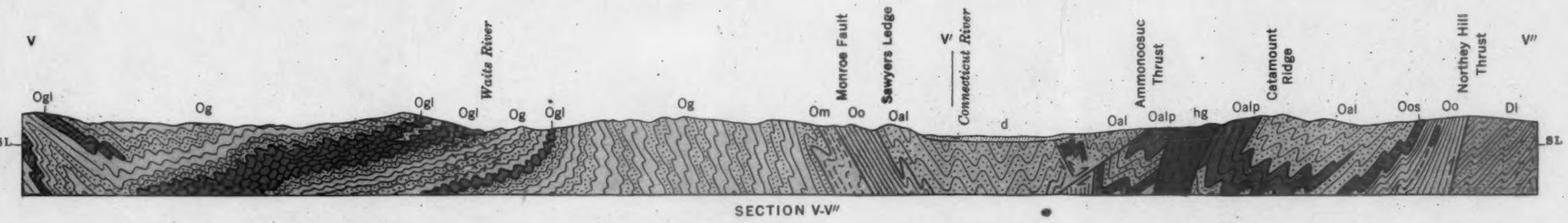
Fault

T denotes overthrust side of thrust faults where known.

Silicified fault zone







GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE WOODSVILLE QUADRANGLE, VERMONT AND NEW HAMPSHIRE

schist. Around the biotite isograd they are albite-epidote amphibolite. West of the garnet isograd they are amphibolite.)

Ihdc

Leighton Hill dike complex

(Numerous dikes of green schist and albite-epidote amphibolite constitute the only outcrops.)

Lower

Dl

Littleton formation

(Mica schist, mica-quartz schist, and micaceous quartzite; biotite and garnet porphyroblasts locally common although in many instances they have been chloritized.)

The lower part of the Littleton formation, all of the Clough formation, all of the Fitch formation, and the upper part of the Partridge formation are missing in this quadrangle because of the Northey Hill thrust.

Upper?

Op

Partridge formation

(Black slate with local thin beds of fine-grained, light-colored quartzite.)

Oam

Ammonoosuc volcanics

(Schistose soda-rhyolite with bluish quartz grains, soda-rhyolite volcanic conglomerate, and chlorite schist; slate is locally abundant especially near the base of the formation one mile north of Woodsville.)

Oal Oalp

Albee formation

(West of the Ammonoosuc thrust is chiefly quartzite, quartzose phyllite, and phyllite. East of the Ammonoosuc thrust the formation consists of white to gray quartzite, micaceous quartzite, mica-quartz schist, and mica schist; the mica schist has porphyroblasts of biotite, garnet and/or staurolite; in a narrow belt north of the French Pond granite the mica schists are coarser grained than the corresponding rocks elsewhere and contain small amounts of sillimanite. The Piermont member, Oalp, consists of rocks similar to those in the main part of the formation but the dark-gray schists with biotite, garnet, and staurolite porphyroblasts are distinctly dominant over the light-colored quartzites and micaceous quartzites.)

ORDOVICIAN?

Middle?

Oos Oo

Orfordville formation

(West of the Ammonoosuc thrust the Orfordville formation consists chiefly of black slate, quartzite, and soda-rhyolite tuff; at the top of the formation the Sunday Mountain volcanics, Oos, consist chiefly of slaty soda-rhyolite tuff; the Hardy Hill quartzite, Ooh, is a thin but distinctive quartzite and quartz conglomerate. East of the Ammonoosuc thrust the Orfordville formation consists chiefly of dark-gray mica schist, with porphyroblasts of biotite and garnet, quartz-mica schist, and thin quartzite; at the top of the formation are the Sunday Mountain volcanics, Oos, consisting of fine-grained amphibolite; the Hardy Hill quartzite, Ooh, is a thin but distinctive quartzite and quartz conglomerate.)

The lower part of the Orfordville formation, present elsewhere, is missing in this quadrangle because of the Monroe fault.

Om

Meetinghouse slate

(Dark mica schist or phyllite, with occasional thin beds of quartzite.)

Og Ogl

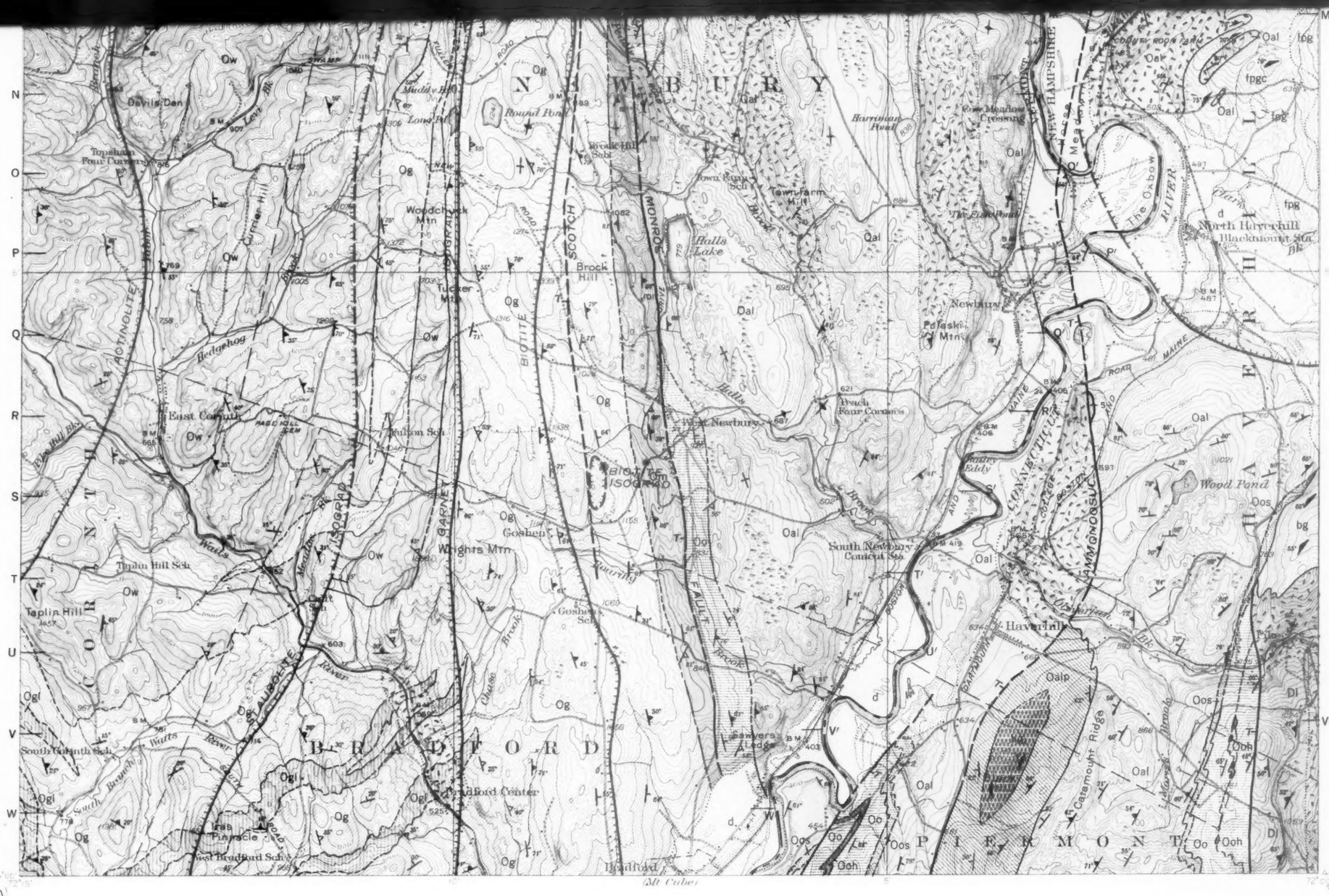
Gile Mountain formation

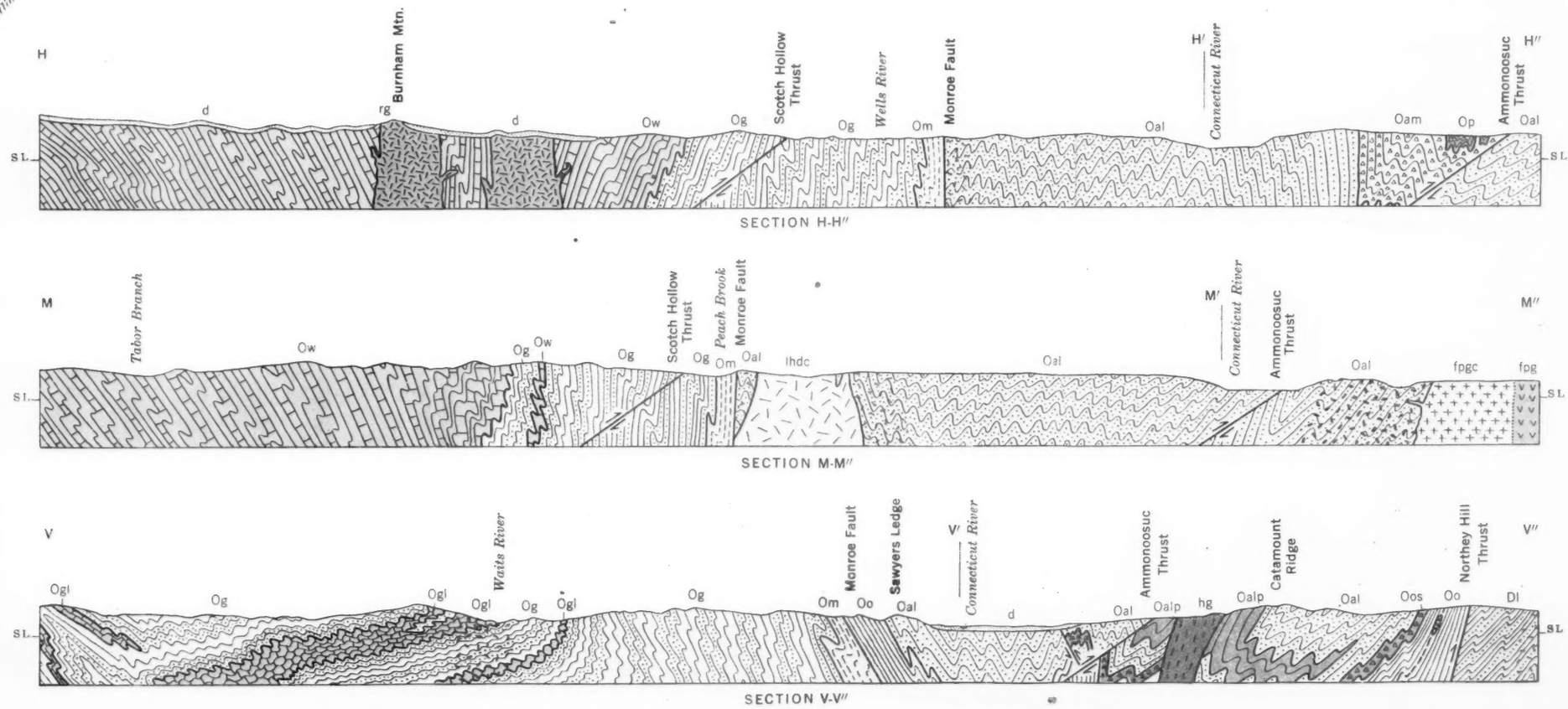
East of the garnet isograd the rocks are phyllite, quartzose phyllite, and micaceous quartzite; west of the garnet isograd the rocks are mica schist, quartz-mica schist, and micaceous quartzite; west of the biotite isograd these rocks contain biotite porphyroblasts, west of the garnet isograd they contain biotite and/or garnet porphyroblasts, and west of the staurolite isograd they contain biotite, garnet and/or staurolite porphyroblasts. Locally calcareous beds are present Ogl.)

Ow

Waits River formation

(Mica schist, quartz-mica schist, calcareous mica schist, and quartzose marble. Many of these rocks contain intermediate to calcic plagioclase, biotite, muscovite, actinolite, clinozoisite, diopside, and garnet.)





GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE WOODSVILLE QUADRANGLE, VERMONT AND NEW HAMPSHIRE

Green schist. Around the biotite isograd they are green schist. Around the biotite isograd they are albite-epidote amphibolite. West of the garnet isograd they are amphibolite.)

Ihdc

Leighton Hill dike complex

(Numerous dikes of green schist and albite-epidote amphibolite constitute the only outcrops.)

Lower

Dl

Littleton formation

(Mica schist, mica-quartz schist, and micaceous quartzite; biotite and garnet porphyroblasts locally common although in many instances they have been chloritized.)

The lower part of the Littleton formation, all of the Clough formation, all of the Fitch formation, and the upper part of the Partridge formation are missing in this quadrangle because of the Northey Hill thrust.

Upper?

Op

Partridge formation

(Black slate with local thin beds of fine-grained, light-colored quartzite.)

Oam

Ammonoosuc volcanics

(Schistose soda-rhyolite with bluish quartz grains, soda-rhyolite volcanic conglomerate, and chlorite schist; slate is locally abundant especially near the base of the formation one mile north of Woodsville.)

Oal Oalp

Albee formation

(West of the Ammonoosuc thrust is chiefly quartzite, quartzose phyllite, and phyllite. East of the Ammonoosuc thrust the formation consists of white to gray quartzite, micaceous quartzite, mica-quartz schist, and mica schist; the mica schist has porphyroblasts of biotite, garnet and/or staurolite; in a narrow belt north of the French Pond granite the mica schists are coarser grained than the corresponding rocks elsewhere and contain small amounts of sillimanite. The Piermont member, Oalp, consists of rocks similar to those in the main part of the formation but the dark-gray schists with biotite, garnet, and staurolite porphyroblasts are distinctly dominant over the light-colored quartzites and micaceous quartzites.)

ORDOVICIAN?

Middle?

Oos Oo Ooh

Orfordville formation

(West of the Ammonoosuc thrust the Orfordville formation consists chiefly of black slate, quartzite, and soda-rhyolite tuff; at the top of the formation the Sunday Mountain volcanics, Oos, consist chiefly of slaty soda-rhyolite tuff; the Hardy Hill quartzite, Ooh, is a thin but distinctive quartzite and quartz conglomerate. East of the Ammonoosuc thrust the Orfordville formation consists chiefly of dark-gray mica schist, with porphyroblasts of biotite and garnet, quartz-mica schist, and thin quartzite; at the top of the formation are the Sunday Mountain volcanics, Oos, consisting of fine-grained amphibolite; the Hardy Hill quartzite, Ooh, is a thin but distinctive quartzite and quartz conglomerate.)

The lower part of the Orfordville formation, present elsewhere, is missing in this quadrangle because of the Monroe fault.

Om

Meetinghouse slate

(Dark mica schist or phyllite, with occasional thin beds of quartzite.)

Og Ogl

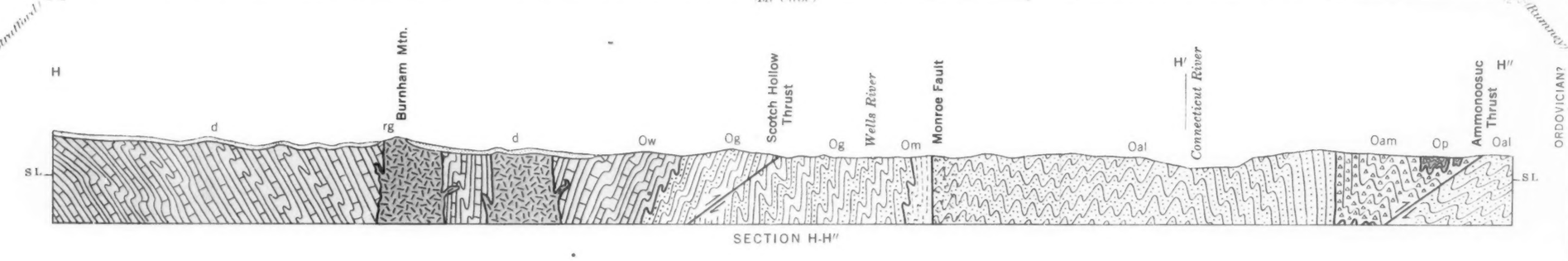
Gile Mountain formation

East of the garnet isograd the rocks are phyllite, quartzose phyllite, and micaceous quartzite; west of the garnet isograd the rocks are mica schist, quartz-mica schist, and micaceous quartzite; west of the biotite isograd these rocks contain biotite porphyroblasts, west of the garnet isograd they contain biotite and/or garnet porphyroblasts, and west of the staurolite isograd they contain biotite, garnet and/or staurolite porphyroblasts. Locally calcareous beds are present Ogl.)

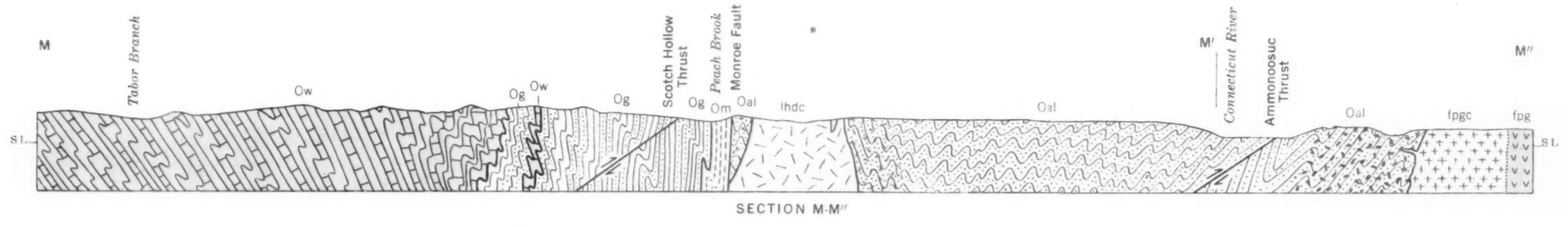
Ow

Waits River formation

(Mica schist, quartz-mica schist, calcareous mica schist, and quartzose marble. Many of these rocks contain intermediate to calcic plagioclase, biotite, muscovite, actinolite, clinozoisite, diopside, and garnet.)



SECTION H-H''



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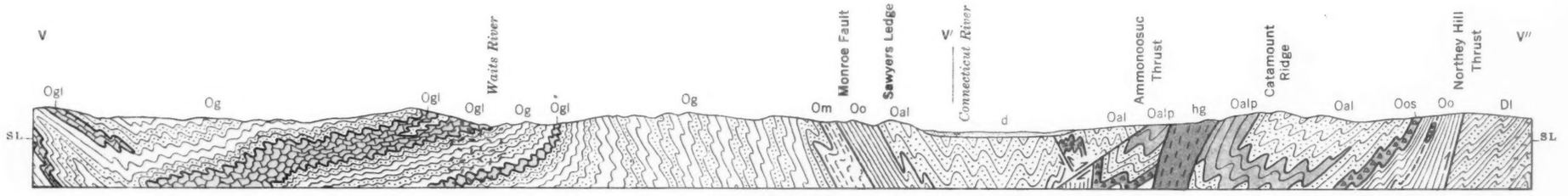


SECTION V-V''

GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE WOODSVILLE QUADRANGLE, VERMONT AND NEW HAMPSHIRE

Topographic base by U. S. Geological Survey. Surveyed in 1931 and 1933-35

Scale 1/62500

1 0 1 2 3 4 Miles

5000 0 5000 10000 15000 20000 Feet

1 0 1 2 3 5 Kilometers

Contour interval 20 feet

Datum is mean sea level

Geology west of Connecticut River by Walter S. White, assisted by E. R. Breed, J. H. Eric, R. Hornblower, Jr., and R.E. Story. Surveyed 1937-39, under direction of M. P. Billings. Geology east of Connecticut River surveyed by Marland P. Billings assisted by S. Bowditch, A. Waldo, and C. W. Wolfe. Surveyed 1934-35.

Survey financed in part by the aid of grants from Harvard University

Published in 1951.

ORDOVICIAN?

Upper?

schist; slate is locally abundant especially near the base of the formation one mile north of Woodsville.)

Oal / Oalp

Albee formation

(West of the Ammonoosuc thrust is chiefly quartzite, quartzose phyllite, and phyllite. East of the Ammonoosuc thrust the formation consists of white to gray quartzite, micaceous quartzite, mica-quartz schist, and mica schist; the mica schist has porphyroblasts of biotite, garnet and/or staurolite; in a narrow belt north of the French Pond granite the mica schists are coarser grained than the corresponding rocks elsewhere and contain small amounts of sillimanite. The Piermont member, Oalp, consists of rocks similar to those in the main part of the formation but the dark-gray schists with biotite, garnet, and staurolite porphyroblasts are distinctly dominant over the light-colored quartzites and micaceous quartzites.)

Middle?

Oos / Oo / Ooh

Orfordville formation

(West of the Ammonoosuc thrust the Orfordville formation consists chiefly of black slate, quartzite, and soda-rhyolite tuff; at the top of the formation the Sunday Mountain volcanics, Oos, consist chiefly of slaty soda-rhyolite tuff; the Hardy Hill quartzite, Ooh, is a thin but distinctive quartzite and quartz conglomerate. East of the Ammonoosuc thrust the Orfordville formation consists chiefly of dark-gray mica schist, with porphyroblasts of biotite and garnet, quartz-mica schist, and thin quartzite; at the top of the formation are the Sunday Mountain volcanics, Oos, consisting of fine-grained amphibolite, the Hardy Hill quartzite, Ooh, is a thin but distinctive quartzite and quartz conglomerate.)

The lower part of the Orfordville formation, present elsewhere, is missing in this quadrangle because of the Monroe fault.

Om

Meetinghouse slate

(Dark mica schist or phyllite, with occasional thin beds of quartzite.)

Og / Ogl

Gile Mountain formation

East of the garnet isograd the rocks are phyllite, quartzose phyllite, and micaceous quartzite; west of the garnet isograd the rocks are mica schist, quartz-mica schist, and micaceous quartzite; west of the biotite isograd these rocks contain biotite porphyroblasts, west of the garnet isograd they contain biotite and/or garnet porphyroblasts, and west of the staurolite isograd they contain biotite, garnet and/or staurolite porphyroblasts. Locally calcareous beds are present Ogl.)

Ow

Waits River formation

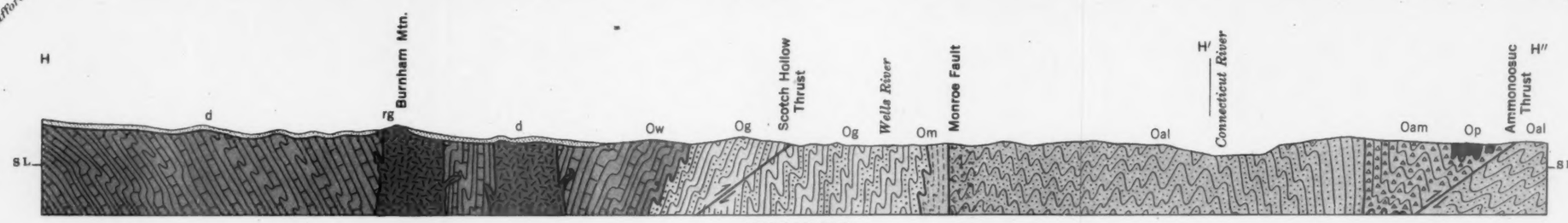
(Mica schist, quartz-mica schist, calcareous mica schist, and quartzose marble. Many of these rocks contain intermediate to calcic plagioclase, biotite, muscovite, actinolite, clinozoisite, diopside, and garnet.)

BED ROCK NOT EXPOSED

d

Glacial drift and alluvium

(Shown only where exceptionally thick and extensive. No attempt has been made on this map to show Quaternary deposits systematically.)



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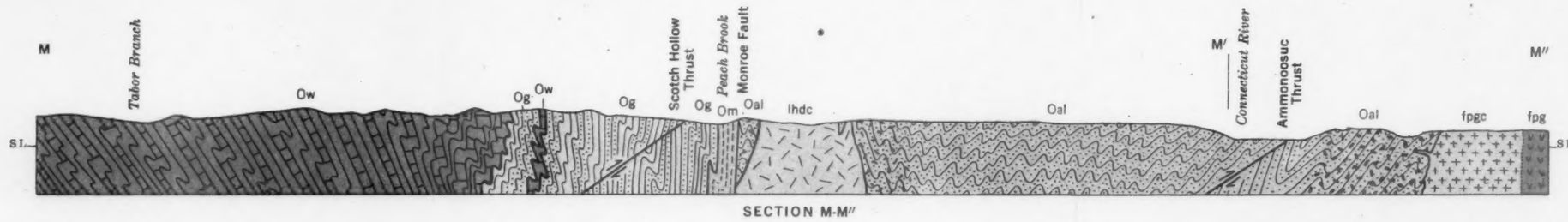

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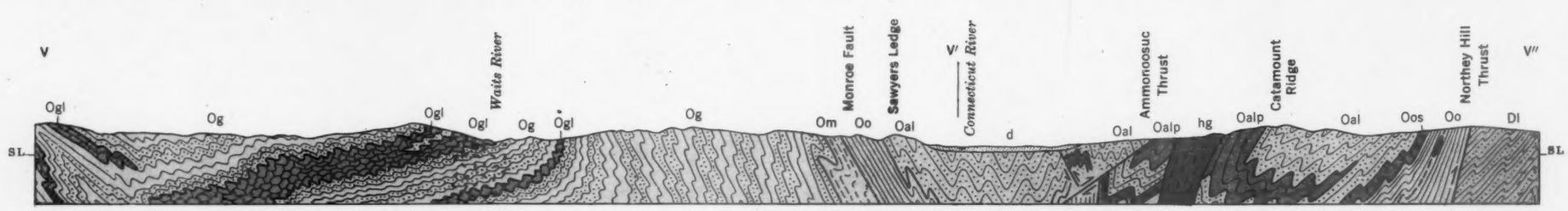

SECTION V-V''

GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE WOODSVILLE QUADRANGLE, VERMONT AND NEW HAMPSHIRE

Topographic base by U. S. Geological Survey. Surveyed in 1931 and 1933-35

Scale 1/62500

1 0 1 2 3 4 Miles
5000 0 5000 10000 15000 20000 Feet
1 0 1 2 3 4 5 Kilometers

Contour interval 20 feet

Datum is mean sea level

Geology west of Connecticut River by Walter S. White, assisted by E. R. Breed, J. H. Eric, R. Hornblower, Jr., and R.E. Story. Surveyed 1937-39, under direction of M. P. Billings. Geology east of Connecticut River surveyed by Marland P. Billings assisted by S. Bowditch, A. Waldo, and C. W. Wolfe. Surveyed 1934-35.
Survey financed in part by the aid of grants from Harvard University
Published in 1951.

ORDOVICIAN?

Upper?

schist; slate is locally abundant especially near the base of the formation one mile north of Woodsville.)

Oal

Albee formation

(West of the Ammonoosuc thrust is chiefly quartzite, quartzose phyllite, and phyllite. East of the Ammonoosuc thrust the formation consists of white to gray quartzite, micaceous quartzite, mica-quartz schist, and mica schist; the mica schist has porphyroblasts of biotite, garnet and/or staurolite; in a narrow belt north of the French Pond granite the mica schists are coarser grained than the corresponding rocks elsewhere and contain small amounts of sillimanite. The Piermont member, Oalp, consists of rocks similar to those in the main part of the formation but the dark-gray schists with biotite, garnet, and staurolite porphyroblasts are distinctly dominant over the light-colored quartzites and micaceous quartzites.)

Middle?

Oos
Oo

Orfordville formation

(West of the Ammonoosuc thrust the Orfordville formation consists chiefly of black slate, quartzite, and soda-rhyolite tuff; at the top of the formation the Sunday Mountain volcanics, Oos, consist chiefly of slaty soda-rhyolite tuff; the Hardy Hill quartzite, Ooh, is a thin but distinctive quartzite and quartz conglomerate. East of the Ammonoosuc thrust the Orfordville formation consists chiefly of dark-gray mica schist, with porphyroblasts of biotite and garnet, quartz-mica schist, and thin quartzite; at the top of the formation are the Sunday Mountain volcanics, Oos, consisting of fine-grained amphibolite; the Hardy Hill quartzite, Ooh, is a thin but distinctive quartzite and quartz conglomerate.)

The lower part of the Orfordville formation, present elsewhere, is missing in this quadrangle because of the Monroe fault.

Om

Meetinghouse slate

(Dark mica schist or phyllite, with occasional thin beds of quartzite.)

Og
Ogl

Gile Mountain formation

East of the garnet isograd the rocks are phyllite, quartzose phyllite, and micaceous quartzite; west of the garnet isograd the rocks are mica schist, quartz-mica schist, and micaceous quartzite; west of the biotite isograd these rocks contain biotite porphyroblasts, west of the garnet isograd they contain biotite and/or garnet porphyroblasts, and west of the staurolite isograd they contain biotite, garnet and/or staurolite porphyroblasts. Locally calcareous beds are present Ogl.)

Ow

Waits River formation

(Mica schist, quartz-mica schist, calcareous mica schist, and quartzose marble. Many of these rocks contain intermediate to calcic plagioclase, biotite, muscovite, actinolite, clinozoisite, diopside, and garnet.)

BED ROCK NOT EXPOSED

d

Glacial drift and alluvium

(Shown only where exceptionally thick and extensive. No attempt has been made on this map to show Quaternary deposits systematically.)

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 647-696, 21 FIGS., 3 PLS. JUNE 1951

GEOLOGY OF THE WOODSVILLE QUADRANGLE, VERMONT-NEW HAMPSHIRE

BY WALTER S. WHITE AND MARLAND P. BILLINGS

ABSTRACT

The Woodsville quadrangle, in east-central Vermont and west-central New Hampshire, was studied primarily to determine the stratigraphic and structural relations between two contrasting sedimentary sequences, one in western and central New Hampshire, the other in eastern and central Vermont. The Vermont rocks in this quadrangle consist of three formations, probably Middle Ordovician, with a total thickness between 15,000 and 20,000 feet. A fault separates these rocks from the New Hampshire rocks, which are pre-Silurian, Silurian, and Lower Devonian and total approximately 20,000 feet thick. Volcanic rocks, mostly pyroclastic, constitute about 10 per cent of the stratified rocks in the Woodsville quadrangle. Plutonic and hypabyssal rocks, probably late Devonian, occurs as dikes, sills, large intrusive sheets, and small stocks.

The quadrangle is divided into four major tectonic units by three faults, which trend north-northeast. The Ammonoosuc fault dips about 36°W. and is younger than the folding and regional metamorphism. The Northey Hill and Monroe faults are steep and are older than most of the metamorphism. Because of the faults the major folds are fragmentary, preserved only as single limbs or as parts of synclines. Countless minor folds are present throughout the region.

Progressive metamorphism is well displayed in the quadrangle. Whereas the rocks in a central north-south belt belong in the chlorite zone, the metamorphism increases toward the east and west margins of the quadrangle, where the rocks lie in the staurolite zone. The sillimanite zone has been attained locally around plutonic rocks. Retrograde metamorphism is especially strong along the Ammonoosuc thrust.

The folding, thrusting, metamorphism, and igneous intrusions were all phases of the late Devonian Acadian orogeny.

Special attention was given to the origin and chronology of the cleavage and minor folds in the Vermont part of the quadrangle, where there is evidence of two stages of deformation. During the earlier stage, schistosity formed parallel to the axial planes of contemporaneous minor folds. Because of the later deformation, these earlier folds are inconspicuous. During the later stage, slip cleavage formed parallel to the axial planes of contemporaneous minor folds. In many places these later folds are very conspicuous and if interpreted as ordinary drag folds may lead to an incorrect interpretation of the stratigraphy and major structure. Toward the west the slip cleavage becomes schistosity. The concept of two stages of deformation has important applications to the geology of New England and other metamorphic terranes.

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Introduction

The Woodsville quadrangle, occupying 210 square miles in east-central Vermont and west-central New Hampshire, lies between 44°00′ and 44°15′ north latitude and 72°00′ and 72°15′ west longitude (Fig. 1). Billings mapped the geology of the New Hampshire part of the quadrangle during parts of the summers of 1934 and 1935, supplemented by brief visits in 1936 and 1937. White mapped the Vermont portion of the quadrangle during the summers

of 1937, 1938, and 1939. In 1940, 1943, and 1946, he spent several months in near-by parts of Vermont in the course of investigations for the U. S. Geological Survey, and this work contributed much to his understanding of the regional setting of the Woodsville area. Billings was assisted during 1934 by Allen Waldo, C. Wroe Wolfe, and Samuel Bowditch. White was assisted for one summer each by E. R. Breed, R. F. Story, J. H. Eric, and Ralph Hornblower, Jr. The Associates in Science of Harvard University helped defray part of the field and laboratory expenses.

Problems

Billings undertook the mapping of the New Hampshire portion of the quadrangle in order that J. B. Hadley, who began mapping the Mt. Cube quadrangle in 1934, might have a better correlation with the stratigraphy of the Littleton-Moosilauke area (Billings, 1937; Hadley, 1942). A topographic map of the Vermont part of the Woodsville quadrangle was not then available, and consequently the mapping ended at the Connecticut River.

During the field work in western New Hampshire between 1931 and 1935, it was recognized that the strata in east-central Vermont differ in many respects from those in west-central New Hampshire. It was tentatively believed that the strata in Vermont underlay those in New Hampshire. The Vermont portion of the Woodsville quadrangle was believed to be a favorable place to study these relations. A topographic map became available about this time and White undertook the investigation. The area was not as satisfactory as hoped, because a fault separates the two contrasting lithological types. On the other hand, since White initiated his study 13 years ago, many additional data bearing on this problem have been obtained in adjacent regions.

The investigation inevitably developed into a study of the stratigraphy and structure of the entire area. The results thus constitute a contribution to a long-range program in central New England to elucidate the paleogeography, structure, metamorphism, and petrology of a complex orogenic region. Moreover, the Vermont portion of the area proved to be a particularly fruitful one in which to study two stages of deformation. The earlier stage was characterized by obscure minor folds and a widespread schistosity that is commonly parallel to the bedding. During the later stage of deformation, the bedding and earlier schistosity were thrown into minor folds associated with a second generation of cleavage.

Stratigraphy

General Statement

The stratified rocks of the Woodsville quadrangle range from Middle Ordovician (?) to Lower Devonian (Table 1). In general, the rocks are progressively younger from west to east, and the youngest rocks are in the extreme southeast corner of the area (Pl. 1). The total thickness of the sedimentary column approximates 35,000 to 40,000 feet, but extensive folding makes it difficult to obtain a precise figure.

Seven formations have been assigned to the Ordovician (?): from oldest to youngest they are the Waits River, Gile Mountain, Meetinghouse, Orfordville, Albee, Ammonoosuc, and Partridge formations. Paleontological evidence indicates that the Waits River is Middle Ordovician and that the Partridge is Pre-Silurian.

Two Silurian formations, the Clough quartzite and Fitch formation, which elsewhere overlie the Partridge formation, are absent in this quadrangle because of the Northey Hill thrust. Part of the Devonian Littleton formation is exposed in the extreme southeast corner of the area.

The Orfordville and the formations above it, although extending a short distance into eastern Vermont (Fig. 2; Pl. 1) are largely confined to New Hampshire; they may be termed the "New Hampshire sequence." The Meetinghouse and the formations west of it, although extending into northern New Hampshire, are far more extensively developed in eastern Vermont; they may be termed the "Vermont sequence." These words are put in quotation marks because it is not intended that they should be used as stratigraphic terms.

Although the structure and metamorphism will be described in detail on later pages, the presentation here of two important facts will greatly facilitate the description of the stratig-

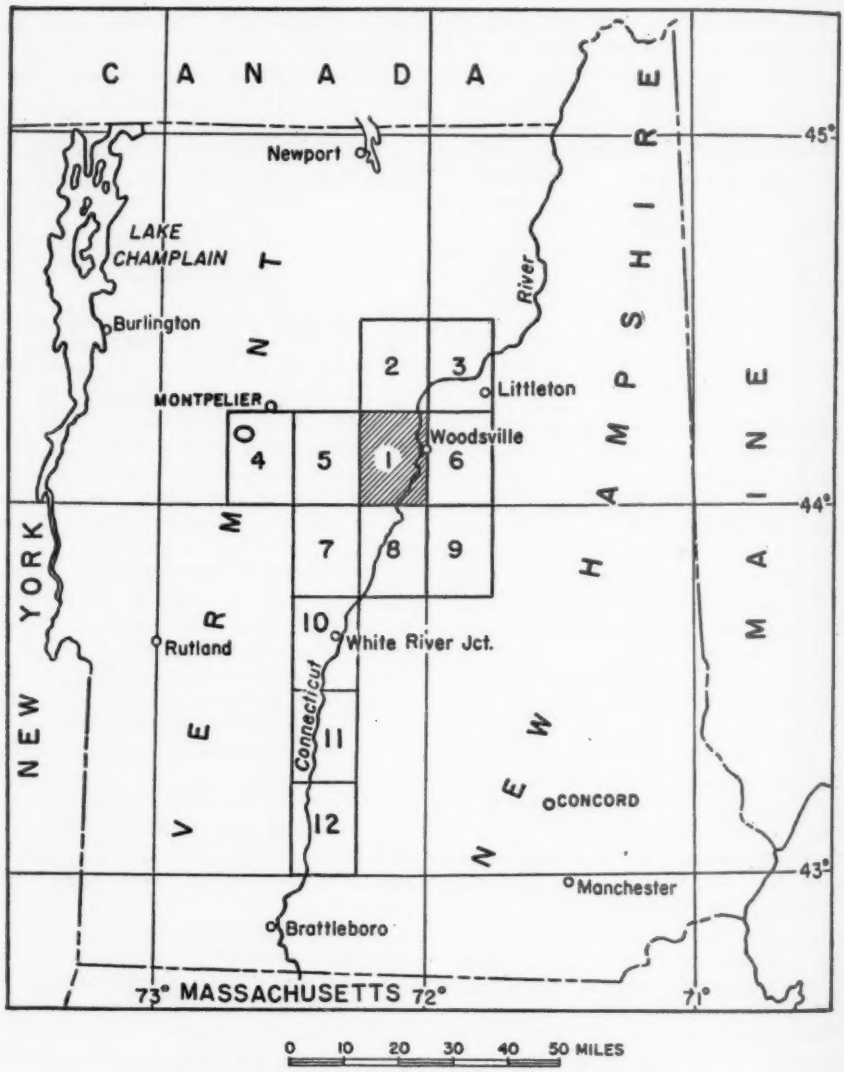


FIGURE 1.—INDEX MAP, VERMONT AND NEW HAMPSHIRE

Woodsville quadrangle is shaded. Quadrangles numbered as follows: 1. Woodsville, 2. St. Johnsbury, 3. Littleton, 4. Barre, 5. East Barre, 6. Moosilauke, 7. Strafford, 8. Mt. Cube, 9. Rumney, 10. Hanover, 11. Claremont, 12. Bellows Falls.

raphy. Three large faults divided the area into four tectonic units; from east to west they are the Northey Hill, Ammonoosuc, and Monroe faults (Fig. 2; Pl. 1). Secondly, the intensity of the metamorphism is not the same throughout the area. A belt of low-grade metamorphism (chlorite zone) extends from the south-central edge of the quadrangle to the northeast corner. The metamorphism increases westward and the biotite, garnet, and staurolite isograds

TABLE 1.—STRATIGRAPHY OF THE WOODSVILLE QUADRANGLE

	Approximate Thickness in feet
Lower Devonian	
Littleton Formation: Gray mica schist, quartz-mica schist, and micaceous quartzite; biotite and garnet porphyroblasts, which in many places are chloritized.	5000
Silurian Clough and Filch formations missing because of Northey Hill thrust	
Upper Ordovician?	
Partridge Formation: Black slate, with some thin beds of fine-grained, light-colored quartzite	0–2000
Ammonoosuc Volcanics: Fine-grained massive to schistose soda-rhyolite, soda-rhyolite volcanic conglomerate, and chlorite schist.	2000
Albee Formation:	
West of Ammonoosuc thrust: Phyllite, feldspathic phyllite, quartzose phyllite, argillaceous quartzite, quartzite, and feldspathic quartzite. Contains biotite porphyroblasts in north-central part of quadrangle.	
East of Ammonoosuc thrust: Mica schist, mica-quartz schist, micaceous quartzite, and quartzite; near French Pond granite contains sillimanite schist. The Piermont member contains rocks similar to main part of formation, but is dominantly dark-gray staurolite-garnet schist, mica-quartz schist, and mica schist.	5000±
Middle Ordovician?	
Orfordville Formation:	
West of Ammonoosuc thrust: Dark-gray to black phyllite with thin beds of gray micaceous quartzite. Sunday Mountain volcanic member at top is gray soda-rhyolite and chlorite schist. Hardy Hill quartzite 1000 feet below top.	
East of Ammonoosuc thrust: Dark-gray to black mica schist and quartz-mica schist, with some thin quartzite beds. Sunday Mountain volcanic member at top is biotite gneiss and amphibolite. Hardy Hill quartzite 700 feet below top.	5000±?
Lower part of Orfordville formation and possibly some unknown formations missing along Monroe fault	
Meetinghouse Slate: Thinly-bedded dark-gray to black slate (southern part of quadrangle) or gray fissile schist (northern part of quadrangle), with some light-gray micaceous quartzite.	2500
Gile Mountain Formation:	
East of biotite isograd: Dark-gray to black phyllite and light-gray micaceous quartzite.	
West of biotite isograd: Dark-gray to black mica schist, mica-quartz schist, gray quartz-mica schist, and light-gray micaceous quartzite. Biotite, garnet, and staurolite porphyroblasts west of appropriate isograd. Also some calcareous beds at south end of quadrangle.	6500
Waits River Formation: Bluish-gray granular quartzose marble, gray calcareous quartz-mica schist, dark-gray to black quartz-mica schist, and mica schist; small amounts of light-gray or buff quartzite and micaceous quartzite.	10,000

appear in orderly succession (Pl. 1). The rocks are classified into metamorphic zones as follows: chlorite zone—east of biotite isograd; biotite zone—between biotite and garnet isograds; garnet zone—between garnet and staurolite isograds; staurolite zone—west of staurolite isograd.The significance of the actinolite isograd (Pl. 1) will be considered more fully later; for the present it will suffice to consider it an isograd of secondary importance lying within the staurolite zone. Metamorphism also increases eastward from the central belt of low-grade rocks, but here some of the metamorphic zones have been eliminated because of thrusting along the Ammonoosuc fault, to the east of which the rocks lie in the staurolite and sillimanite zones. The lithological description of each formation is consequently more complicated than in an area of uniform metamor-

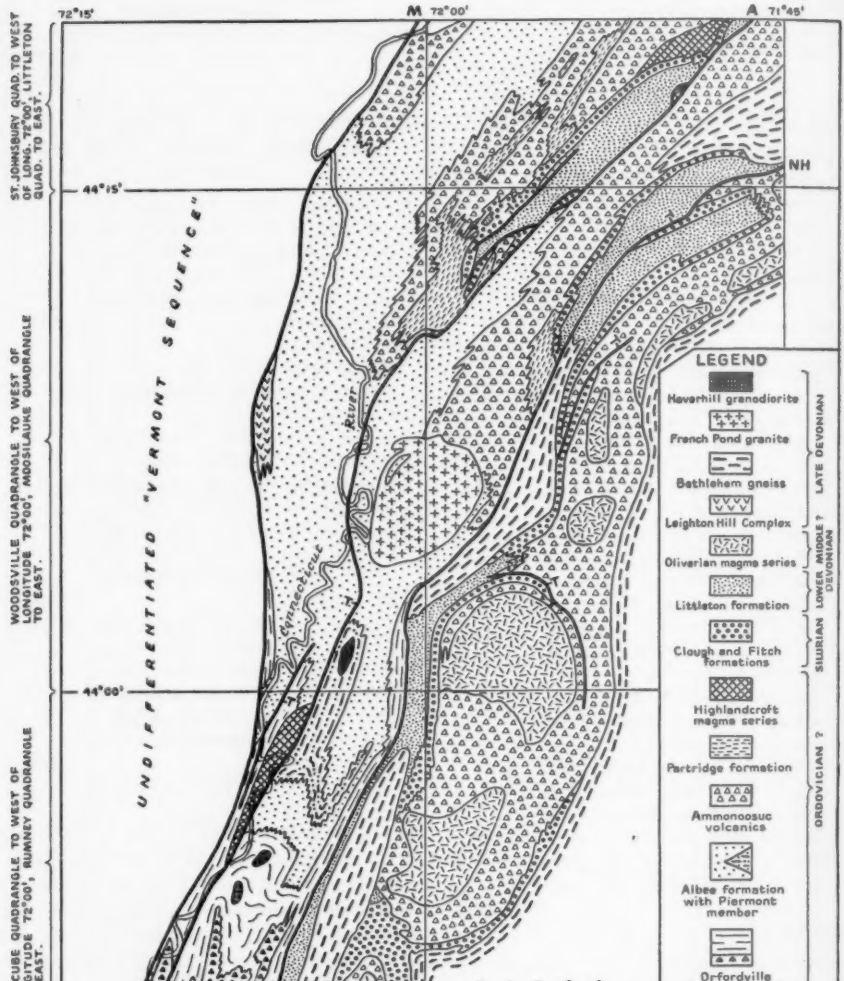


FIGURE 2.—GEOLOGICAL SKETCH MAP OF NEW HAMPSHIRE PART OF WOODSVILLE QUADRANGLE AND ADJACENT AREAS

Faults shown by heavy lines. M = Monroe fault; A = Ammonoosuc thrust; NH = Northey Hill thrust; T = overthrust side.

phism and must be presented separately for each metamorphic zone.

Waits River Formation

Distribution.—The Waits River formation underlies most of the western third of the Woodsville quadrangle (Pl. 1), and forms a belt 2 to 25 miles wide extending from the northern to the southern boundary of Vermont (Hitchcock *et al.*, 1861, map; formation called "calciferous mica schist").

Lithologic character.—The Waits River formation (Currier and Jahns, 1941, p. 1491) is

TABLE 2.—ESTIMATED MODES OF THE WAITS RIVER FORMATION

	1	2	3	4	5	6	7	8	9	10	11	12	13
Number of sections	2	5	5	4	1	1	5	1	2	1	1	2	1
Quartz	76	41	41	32	59	17	27	48	40	10	88	44	70
Plagioclase	1	x	5	x	5	50	6	15	14	30			10
Carbonate	3	54	31	46			40		15	10			t
Muscovite	6	2	4							30	x	35	2
Biotite	12	2	18	10	3	20		8		15	10	17	15
Chlorite		x	x								1	1	3
Actinolite				12*	5	10	3	10	7*				
Clinozoisite					8	2	7	2	12	5			
Diopside							17	2	8				
Garnet									4		1	1	
Microcline					15		x	12					
Tourmaline		t	t		x		t				x	t	x
Sphene		t	t	t	x	1	x	1	x		x	t	x
Apatite			t					x			x	t	
Zircon		t		t	x		t	x	t		x		x
Carbon	2	1	1	x	3		t	1				2	
Metallic opaque		x	x	t	2	x	x	1	t		x	x	x
Per cent of anorthite in plagioclase	50	30	30–60	35–50	46	50	29–95	35	40–90	66			20

* Hornblende in one section

Grain size: Matrix, 0.1–0.5 mm.; porphyroblasts, 0.5–4.0 mm.

x—present in 50 per cent or more of thin sections

t—present in less than 50 per cent of thin sections

Blank—not seen in any of thin sections

1. Calcareous micaceous quartzite
2. Quartzose magnesian marble
3. Calcareous mica schist
4. Lime-silicate rock
5. Lime-silicate rock
6. Lime-silicate rock
7. Lime-silicate rock
8. Lime-silicate rock
9. Lime-silicate rock
10. Lime-silicate rock
11. Quartzite
12. Quartz-mica schist
13. Feldspathic quartzite

made up of interbedded calcareous and non-calcareous schists. The calcareous rocks, which constitute from 40 to 50 per cent of the whole formation, range from bluish-gray granular quartzose marble to gray calcareous quartz-mica schist. These rocks weather dark brown. The noncalcareous rocks are dominantly dark-gray to black quartz-mica schist and mica schist, with subordinate light-gray or buff micaceous quartzite and quartzite. The calcareous rocks are interbedded with the non-calcareous rocks in all proportions; some sections several hundred feet thick are dominantly calcareous, whereas others have little or no carbonate. Individual beds of either calcareous or non-calcareous rock range from a fraction of an inch to over 10 feet thick; the distinctive beds of quartzose marble are typically 1 inch to 2 feet thick.

Over most of the area, the dominant minerals in the formation are quartz, calcite, dolomite, biotite, and muscovite (Table 2). The calcareous rocks locally contain porphyroblasts of actinolite in the western part of the quadrangle, and diopside, clinozoisite, and anorthite are prominent porphyroblasts in highly metamorphosed marbles immediately adjacent to granodiorite in the northwest corner of the area. The lime-silicate rocks contain little or no dolomite, but the ratio of calcite to dolomite was not determined in the rocks that contain both.

The rocks are all medium- to coarse-grained

and generally schistose. Schistosity is best developed in the more micaceous rocks, in which microscopic layers of muscovite and biotite separate discontinuous layers or lenses of quartz or quartz and calcite. Dimensional elongation of all minerals parallel to the banding is very good, but the crystallographic orientation of scattered porphyroblasts of actinolite and mica is practically random. In the more massive quartzose marble, schistosity is defined by a rude orientation of scattered mica plates and by dimensional elongation of calcite and some quartz grains.

Thickness.—Extreme structural complication within the area underlain by the Waits River formation, and the fact that only one of its stratigraphic boundaries is exposed here, make a reliable estimate of its thickness impossible. In the Barre quadrangle (Fig. 1), where the entire formation is exposed as a simple homoclinal belt, the distance between the upper and lower boundaries, measured normal to bedding, is at least 20,000 feet. Because of repetition of beds in minor folds, this distance may be as much as twice the sum of the present thicknesses of all the beds in the formation (Billings, 1937, p. 475; Currier and Jahns, 1941, p. 1503). The present thickness of the formation, corrected thus for folding, is of the order of 10,000 feet. This present thickness may be considerably less than the thickness at the time of deposition because of thinning of individual beds by rock flowage and shearing during deformation. The amount of such thinning, however, cannot be readily calculated here because of lack of fossils, amygdules, oölites, or other bodies whose original shapes are known.

Correlation.—The calcareous rocks of the western part of the Woodsville quadrangle are correlated with the Waits River formation of the Barre quadrangle (Currier and Jahns, 1941, p. 1491) primarily on the basis of lithologic similarity. The area between the Woodsville and Barre quadrangles is largely underlain by similar rocks, as shown by detailed geologic mapping in the East Barre and Strafford quadrangles (White and Eric, 1944; Doll, 1945; White and Jahns, 1950).

The Waits River formation is younger than the fossiliferous Shaw Mountain formation, which is no older than Middle Ordovician (Currier and Jahns, 1941, p. 1500–1502). Preliminary identification of recently discovered fossils from the Waits River formation near Montpelier (Fig. 1) suggests that it is Middle Ordovician (Cady, 1950). The Waits River formation continues northward into Canada, where it is called the Tomifobia formation and considered to be Middle Ordovician (Clark, 1934, p. 11).

Gile Mountain Formation

Distribution.—The Gile Mountain formation underlies a belt 2 or more miles wide that extends from the southwest corner of the quadrangle to the north-central part. Both the western contact with the Waits River formation and the eastern contact with the Meetinghouse slate are gradational.

Lithologic character.—The Gile Mountain formation is composed of thinly to thickly interbedded dark-gray to black schist (phyllite near the eastern border of the outcrop area) and light-gray micaceous quartzite, either of which may be locally dominant. The schist ranges from almost pure mica schist, with subordinate quartz, to quartz-mica schist. All gradations are found, furthermore, between quartz-mica schist and micaceous quartzite. Feldspathic rocks are not abundant and pure quartzite is rare. Typical outcrops of the formation reveal alternations of micaceous and quartzose beds. Where schist predominates, individual beds are almost everywhere less than 6 inches thick, and are commonly an inch or less thick. Where micaceous quartzite is the most abundant rock, it may occur in beds 3 or 4 feet thick separated by thin schist laminae.

Quartz and muscovite are the principal constituents of the formation as a whole, and small amounts of plagioclase and chlorite are present in most rocks (Table 3). Biotite, garnet, and staurolite are also common in rocks of appropriate metamorphic grade; the degree of metamorphism increases westward, and garnet, for example, is a common minor constituent of the rocks west of the garnet isograd (Pl. 1).

All the rocks of the Gile Mountain formation are schistose. In mica schist and phyllite, quartz and feldspar occur as isolated tabular grains and lenticular aggregates between al-

TABLE 3.—ESTIMATED MODES OF THE GILE MOUNTAIN FORMATION

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Number of sections	8	4	9	2	6	5	3	2	6	3	5	4	2	1	2
Quartz	63	21	64	82	51	22	46	81	51	18	60	84	25	8	50
Plagioclase	2	1	3	1	24	1	2	3	20				11	20	
Muscovite	29	73	20	11	8	58	32	7	7	40	14	1	25	1	3
Chlorite	6	1	4	x	5	8	8	x	1	4	5	1			
Biotite		4	9	6	11	7	10	7	19	24	17	11	30		
Garnet						4	2	1	2	5	2	1			
Staurolite										9	2	2	8		
Sillimanite													1		
Carbonate	t		t		1			1						70	47
Epidote				x	t				t						
Sphene	x	t	x		t	t	x	x	x	x	x	x	x	x	x
Apatite	t	t	x	x	x	t	x	x	x	x	x	x	x	x	t
Zircon	x	t	t	x	x	t		t	t		t	t	x		
Tourmaline	x	x	x	x	x	x	x	x	x	x	x	x	x	x	x
Carbon	x	x	x		t	x	x	x	x	x	x	x	x	1	x
Metallic opaque	x		x	x	x	t	t	t	x		t	t	t	x	x
Per cent of anorthite in plagioclase	10	10	10–35	20	10	20	20–35	5–35	5–30				20–35	60	
Average grain size quartz (mm.)	0.1	0.07	0.2	0.1	0.2	0.05	0.1	0.2	0.2	0.3	0.3	0.2	0.3	0.3	0.4

x—present in 50 per cent or more of thin sections
t—present in less than 50 per cent of thin sections
Blank—not seen in any of thin sections

1. Phyllitic quartzite, chlorite zone
2. Phyllite, biotite zone
3. Quartzose phyllite, biotite zone
4. Micaceous quartzite, biotite zone
5. Feldspathic phyllite, biotite zone
6. Mica schist, garnet zone
7. Quartz-mica schist, garnet zone
8. Biotitic quartzite, garnet zone
9. Feldspathic schist, garnet zone
10. Mica schist, staurolite zone
11. Quartz-mica schist, staurolite zone
12. Biotitic quartzite, staurolite zone
13. Mica schist with sillimanite
14. Feldspathic marble, garnet zone
15. Calcareous quartzite, garnet zone

most perfectly oriented plates of muscovite and biotite. The quartz-mica schists are typically banded, with laminae of parallel mica plates separating thin layers or lenses of quartz or quartz and feldspar. In most rocks, this lamination is parallel to bedding, but inasmuch as it clearly cuts bedding in some places, it is considered to be of tectonic origin. In the micaceous quartzites, muscovite and biotite occur as isolated plates, oriented parallel to schistosity, in a granular matrix of quartz and subordinate feldspar; quartz grains in these rocks are much less tabular, as a rule, than quartz grains in the more micaceous rocks.

Garnet, staurolite, and part of the muscovite and biotite form randomly oriented porphyroblasts in the rocks in which they occur. Typically these minerals have replaced the rock bodily; micaceous laminae adjacent to these porphyroblasts are clearly transected rather than bulged out around them.

At the south end of the Woodsville quadrangle, west of Bradford Center, there are three lenses a few hundred feet thick in which 10 to 50 per cent of the rock is quartzose marble and calcareous schist (Pl. 1). Lithologically these calcareous rocks are essentially the same as those in the Waits River formation; as stratigraphic units, they differ from the Waits River in that a high proportion of the inter-

bedded non-calcareous rock is micaceous quartzite. In Plate 1, a narrow tongue of Gile Mountain is shown west of the main body of the formation in the central part of the quadrangle. Whether this represents an attenuated fold, a fault slice, or a stratigraphic tongue is not clear.

Thickness.—The present thickness of the Gile Mountain formation is best measured near Goshen (Pl. 1), where the displacement on the Scotch Hollow thrust is believed to be small, and where the calculation is complicated by a minimum number of minor folds. The beds here are almost vertical, and the breath of outcrop is about 10,000 feet. The sum of the present thicknesses of all the beds in the formation is probably of the order of 6000 or 7000 feet, using a divisor of 1.5 to correct for repetition of beds in minor folds. This divisor is used because minor folds are rare in the easternmost part of the formation, and in the westernmost part the number of minor folds per unit area is comparable to the number in regions in which a divisor of 2 has seemed reasonable (Billings, 1937, p. 475; Currier and Jahns, 1941, p. 1503). As in the case of the Waits River formation, the amount of thinning of beds because of flowage during deformation cannot be calculated, so the original thickness at the time of deposition is not known.

Correlation.—The Gile Mountain formation has been traced by detailed mapping from the Woodsville quadrangle to Gile Mountain in the Strafford quadrangle (White and Eric, 1944; Hadley, 1950; Doll, 1945). The formation, as defined here, includes the rocks at the type locality and most of the rocks mapped by Doll as "Gile Mountain Schist" (Doll, 1945, p. 18–19) in the remainder of the Strafford quadrangle. It does not, however, include any of the rocks exposed in the southeast corner of the Strafford quadrangle east of the Ammonoosuc thrust that Doll mapped as Gile Mountain; these rocks are areally continuous along their strike with the Orfordville formation of the Mt. Cube quadrangle (Hadley, 1942; 1950, geologic maps). The formation is tentatively assigned to the Middle Ordovician.

Meetinghouse Slate

Distribution.—The Meetinghouse slate (Doll, 1945, p. 19) forms a narrow band that extends northward from the south-central edge of the quadrangle. It gradually narrows northward and disappears about a mile south of the north edge of the quadrangle.

Lithologic character.—The Meetinghouse slate is composed of thinly bedded dark-gray to black slate (southern part of quadrangle) or gray fissile schist (northern part), with very minor amounts of thinly interbedded light-gray micaceous quartzite. The slate or schist is dominantly micaceous, and quartz-mica slate or schist is subordinate. Lithologically the various rock types are indistinguishable in hand specimen from their counterparts in the Gile Mountain formation, and the two formations are separated entirely on the basis of the relative proportions of the micaceous and quartzose rocks.

The Meetinghouse typically contains about 10 per cent of arenaceous phyllite and micaceous quartzite, in beds an inch or less thick, interstratified with the darker argillaceous layers. The proportion of the arenaceous layers increases westward to about 40 per cent at the arbitrary western boundary of the formation, and the thickness of these beds increases correspondingly to as much as a foot.

Thickness.—An unknown amount of the Meetinghouse slate is cut out by the Monroe fault, so the total thickness is not known. The breadth of outcrop near Bradford, where the rocks dip very steeply, is about 2500 feet. This probably represents closely the sum of the present thicknesses of individual beds exposed here, as minor folds are uncommon.

Correlation.—The slate belt has been traced southward by detailed mapping to the type locality in the Strafford quadrangle (Hadley, 1950; Doll, 1945). It is tentatively assigned to the Middle Ordovician.

Orfordville Formation

Distribution.—The Orfordville formation is exposed: (1) just west of the Connecticut River at the south end of the quadrangle; (2) just east of the Connecticut River in the south end of the quadrangle; and (3) in a belt 4 miles long and 0.5 mile wide in the southeast corner of the quadrangle.

Lithologic character.—West of the Connecticut River, the Orfordville is dominantly

dark-gray to black phyllite with thin beds of micaceous quartzite. Seventy feet west of the contact with the Albee formation, a bed of pebble conglomerate is 1 to 2 feet thick; the matrix is arenaceous, whereas the pebbles are quartzite, impure limestone, and phyllite. Two thousand feet south of Sawyers Ledge, both west of the highway and in an island in the Connecticut River, metamorphosed soda-rhyolite tuffs that may represent the Sunday Mountain volcanics have not been separately designated on the geological map (Pl. 1).

Just east of the Connecticut River in the south end of the quadrangle, the Orfordville is chiefly dark-gray to black slate, with some interbedded quartzite and slaty soda-rhyolite. The Hardy Hill quartzite member, a greenish-gray quartzite and quartz conglomerate, is 60 feet or less thick and lies about 1000 feet stratigraphically below the top of the formation. The Sunday Mountain volcanic member, composed of gray soda-rhyolite tuff and chlorite schist, is at the top of the formation and about 200 feet thick. An unusually good section of the Orfordville exposed along the southern border of the quadrangle is given in Table 4. The thicknesses are only approximate, as it is difficult to evaluate with precision the amount of repetition due to folding.

In the southeast corner of the quadrangle, the Orfordville lies in the staurolite zone. The main part of the formation is dark-gray to black mica schist and quartz-mica schist, with some thin quartzite beds. The Hardy Hill quartzite member occurs about 700 feet below the top of the formation. The Sunday Mountain volcanic member, at the top, is about 300 feet thick and is composed of interbedded fine-grained biotite gneiss and amphibolite; these rocks were originally volcanic tuffs and breccias with the composition of soda-rhyolite, andesite, and basalt. A stratigraphic section of this area is given in Table 5.

Detailed petrographic descriptions are given by Hadley (1942, p. 119–123).

Thickness.—Tables 4 and 5 indicate that about 1500 feet of the Orfordville formation are exposed in the Woodsville quadrangle. In the Mt. Cube quadrangle, where lower parts of the formation are exposed, Hadley (1942, p. 123) estimated a thickness of 5000 feet with a possible error of 30 per cent; the base of the formation was not exposed there.

TABLE 4.—STRATIGRAPHIC SECTION OF ORFORDVILLE FORMATION, WESTERN PART OF TOWNSHIP OF PIERMONT

Top of section is Dartmouth College Road at border of Woodsville and Mt. Cube quadrangles

		Thickness in feet (approximate only)
Albee formation	Greenish-gray slaty quartzite	2000
Orfordville formation	Tuff composed of gray slaty soda-rhyolite matrix, with pebbles of soda-rhyolite 1/2 inch long (*Sunday Mountain volcanic member*)	200
	Gray slate and quartzite	200
	Slaty soda-rhyolite tuff; quartz grains 1/8 inch across in some beds	100
	Gray slate and quartzite	225
	Thin-bedded black slate and quartzite	100
	Gap	200
	Calcareous green schist	30
	Quartz conglomerate (*Hardy Hill quartzite member*)	30
	Black slate	400
	Soda-rhyolite tuff; quartz grains 1/8 inch across	5
	Black slate	10

Base of section is edge of sand plain, 1/3 mile south of B.M. 454, western part of township of Piermont.

Correlation.—The rocks assigned to the Orfordville in the southeast corner of the quadrangle may be traced southwestward into the type locality in the Mt. Cube quadrangle (Fig. 2). The rocks in the vicinity of the Connecticut River in the southern part of the quadrangle are correlated with the Orfordville formation for three reasons: (1) they are stratigraphically beneath the Albee formation, as in the type locality, (2) prior to metamorphism, they were very

similar lithologically to those in the type locality, and (3) the Sunday Mountain and Hardy Hill members appear at the appropriate stratigraphic position. This formation is unfossiliferous and in recent years has been tentatively assigned to the Middle Ordovician (Hadley, 1942, p. 124).

TABLE 5.—GENERALIZED STRATIGRAPHIC SECTION OF ORFORDVILLE FORMATION 1.5 MILES SOUTH OF PIKE

		Thickness in feet (approximate only)
Albee formation	Quartz-mica schist with porphyroblasts of biotite, garnet, and some staurolite	
Orfordville formation	Fine-grained amphibolite (*Sunday Mountain volcanic member*)	200
	Dark mica schist with porphyroblasts of biotite and garnet	500
	Quartz conglomerate (*Hardy Hill quartzite member*)	0–100
	Dark mica schist with porphyroblasts of biotite and garnet; actinolite schist, 5 feet thick, 600 feet below top of this member	800
Base of section is Northey Hill thrust		

Albee Formation

Distribution.—The Albee formation occupies much of the eastern half of the quadrangle. Because of a considerable difference in the grade of metamorphism a distinction should be made between a western and an eastern belt. The western belt, lying west of the Ammonoosuc thrust and averaging 4 miles in width, extends from the northeast corner of the quadrangle in a southerly direction to the south end of the quadrangle, where it splits into two tongues that die out a short distance to the south in the Mt. Cube quadrangle (Fig. 2). The eastern belt, east of the Ammonoosuc thrust, extends along the east-central and southeast margins of the quadrangle.

Lithologic character.—The Albee formation west of the Ammonoosuc thrust is in the chlorite zone, except for a long narrow triangular area in the northern part of the quadrangle lying west of the biotite isograd and east of the Monroe fault.

The rocks in the chlorite zone are chiefly phyllite, feldspathic phyllite, quartzose phyllite, argillaceous quartzite, quartzite, and feldspathic quartzite, all six types being about equally abundant. The yellowish-green to bluish-green color of many of these rocks distinguishes this formation from others in the quadrangle.

The phyllites are black, dark-green, or greenish-yellow schistose rocks composed chiefly of quartz, sericite, and chlorite (Table 6, column 1). A variety rich in albite-oligoclase is called feldspathic phyllite (Table 6, column 2). The terms phyllite and feldspathic phyllite are reserved for rocks containing 60 per cent or less of quartz or quartz plus feldspar (Billings, 1937, p. 475). Numerous outcrops are composed exclusively of phyllites or feldspathic phyllites in which the cleavage is well developed but the bedding is difficult or impossible to identify. Microscopic study shows that the sericite and chlorite occur either as small well-oriented, irregular flakes or as matted aggregates in bands parallel to the schistosity. Quartz and feldspar are in small grains scattered through the rock or are concentrated in thin laminae parallel to the schistosity.

Rocks containing 60 to 80 per cent quartz or quartz plus feldspar are termed quartzose phyllites if they posses a good schistosity but are called argillaceous quartzites if they are relatively massive. Such rocks are generally yellowish-green. They may be the only rock present in the outcrop or they may be interbedded with phyllite or quartzite. An average mode is given in Column 3 of Table 6. Some specimens are rich in feldspar, but modes of such rocks have not been prepared.

The quartzites are massive nonfoliated or weakly foliated yellowish-green, buff, or white rocks that contain 80 per cent or more quartz (Table 6, column 4). The feldspathic quartzites contain at least 10 per cent feldspar, the total quartz and feldspar constituting at least 80 per

TABLE 6.—ESTIMATED MODES OF ALBEE FORMATION

	Chlorite Zone							Biotite Zone				Staurolite Zone					Sillimanite Zone	
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
Number of sections	10	6	2	10	4	1	2	7	2	3	1	2	1	2	1	1	3	1
Quartz	34	24	68	84	63	44	60	45	67	85	78	39	40	72	69	84	45	52
Plagioclase		36		3	23		8	t		1	15		25	5	15	5	9	30
Orthoclase						5												
Muscovite*	55	7	21	9	7	25	7	29	29			30	10	10	7	3	22	
Chlorite	10	28	10	3	6	20	18	15	1	8	3	x	3	7	6	1	1	
Epidote†	x	1	t			4	4	1		1	2							x
Carbonate	1	3	1	1	x			1										
Biotite								8	2	5	1	21	17	2		5	18	2
Garnet												2		x		1	x	4
Staurolite												6	5				x	
Sillimanite																	3	
Magnetite		x	x	x	1	2	2	x	1	x			x	2	3	1	2	
Ilmenite											x	x						
Pyrite						x	1	t			1	2		1				
Rutile						x												
Leucoxene				t	t													
Zircon		x		t				t		x				x				
Apatite		t		t	x	x		x	x	t	x	x	x	1		x		
Tourmaline	t	t	x	x	x		x	t	x			x		x			t	
Sphene	t	1		x	x			1	x	x								
Hornblende																		12
Carbon	t		x							t								
Per cent of anorthite in plagioclase		10		8	10		10				10		40	30	40	20	27	80

* Sericite in chlorite zone

† Includes clinozoisite

x—present in 50 per cent or more of thin sections

t—present in less than 50 per cent of thin sections

Blank—not seen in any of thin sections

1. Phyllite
2. Feldspathic phyllite
3. Quartzose phyllite and argillaceous quartzite
4. Quartzite
5. Feldspathic quartzite
6. Hornfels; Oliverian Brook, 3/4 mile north of Village of Haverhill.
7. Hornfels with porphyroblasts of chlorite retrogressed from biotite; Oliverian Brook, 3/4 mile north of Village of Haverhill.
8. Phyllite with biotite porphyroblasts
9. Mica-quartz schist with biotite porphyroblasts
10. Quartzite with biotite porphyroblasts
11. Feldspathic quartzite with biotite porphyroblasts
12. Staurolite-garnet-mica schist
13. Feldspathic staurolite-mica schist
14. Mica-quartz schist
15. Feldspathic mica-quartz schist
16. Quartzite
17. Sillimanite schist
18. Hornblende-bytownite schist

cent of the rock (Table 6, column 5). The quartzites are in beds an inch to 6 feet thick interbedded with phyllites, quartzose phyllites, and argillaceous quartzites.

In Oliverian Brook, 0.8 mile north-northeast of the village of Haverhill, the Albee has been converted to hornfels (Table 6, columns 6 and 7). In the rock listed in column 7 half the chlo-

rite is in porphyroblasts from 0.4 to 0.6 mm. in diameter that contain numerous needles of rutile. This suggests that the chlorite is altered biotite.

The rocks of the Albee formation (Table 6, columns 8–11) lying west of the biotite isograd and east of the Monroe fault are similar to those described in the preceding paragraphs, except for the presence of biotite porphyroblasts 0.1 to 2 mm. in diameter.

The Albee is more highly metamorphosed east of the Ammonoosuc thrust than to the west. In general, the rocks lie in the staurolite zone and such minerals as biotite, garnet, and staurolite are present. Locally, around the French Pond granite, the rocks are in the sillimanite zone. In a belt nearly 1000 feet wide directly east of the Ammonoosuc thrust the rocks have undergone striking retrograde effects, described in the section on metamorphism.

The Albee formation in the staurolite zone consists of about equal amounts of mica schist, micaceous quartzite, and quartzite. Most of these rocks are white to light gray, rather than yellowish-green as in the lower metamorphic zones. Dark-gray types are distinctly subordinate in the main part of the formation.

The mica schists are light-gray to dark-gray schistose rocks with less than 60 per cent quartz or quartz plus feldspar. Most of these rocks have porphyroblasts of garnet or staurolite or both, and some varieties are feldspathic (Table 6, columns 12 and 13). The mica-quartz schists are light-gray schistose rocks with 60 to 80 per cent quartz or quartz plus feldspar (Table 6, columns 14 and 15). The quartzites, which contain over 80 per cent quartz or quartz plus feldspar, are white to gray massive rocks (Table 6, column 16). Some of the quartzites are crossed by narrow biotite-rich bands(Hadley, 1942, Pl. 3), and in many outcrops it is difficult to decide whether these bands are original bedding or fracture cleavage along which biotite has concentrated.

The Albee along the northwest margin of the French Pond granite and in the large inclusion in the granite lies in the sillimanite zone. The mica schists here are coarser-grained than corresponding rocks elsewhere in the Albee formation; muscovite flakes are not uncommonly 1 to 2 mm. in diameter. Sillimanite is not megascopically visible, but thin sections show that small needles 0.1 to 1 mm. long constitute 2 to 4 per cent of the rocks (Table 6, column 17). Modes of the associated quartzites and mica-quartz schists have not been determined.

The Piermont member of the Albee formation underlies an area of about 2 square miles south of the village of Haverhill. It consists of rocks identical with those found in the main part of the Albee formation east of the Ammonoosuc thrust, but the proportions are different. The main part of the Albee formation consists chiefly of light-colored rocks, notably quartzite, micaceous quartzite, and mica-quartz schist, with dark-gray staurolite-garnet schist, mica-quartz schist, and mica schist distinctly subordinate. In the Piermont member, the dark-gray types with biotite, garnet, and staurolite porphyroblasts are dominant. The garnet porphyroblasts are small, pink, euhedral crystals that seldom exceed a millimeter is diameter. The staurolite porphyroblasts, from 0.5 to 3 centimeters long, are euhedral, with the prism well developed but the terminal facies lacking; there are some cruciform twins. Although some staurolite crystals are solidly built, others are mere skeletons and include many other minerals poikilitically. The staurolite crystals are diversely oriented and it is apparent that the crystals formed in large part after the deformation had ceased.

Thickness.—Hadley (1942, p. 146) estimates the Albee formation to be about 5000 feet thick, but recognizes that an accurate calculation is impossible because of the intense minor folding.

Correlation.—The Albee formation west of the Ammonoosuc thrust may be traced continuously from the type locality in the Littleton quadrangle (Fig. 2). The belt east of the Ammonoosuc thrust is contiguous to rocks in the Moosilauke quadrangle that have been correlated with the Albee formation because of their lithologic characteristics and stratigraphic position (Fig. 2). The Albee formation is pre-Silurian and in recent years has been tentatively assigned to the Upper Ordovician (Billings, 1937, p. 475). The age is discussed more fully later.

Ammonoosuc Volcanics

The Ammonoosuc volcanics occupy a belt extending northeasterly from Woodsville. The

contact with the Albee formation is not always easy to place, as there is a transition zone several hundred feet thick in which Albee and Ammonoosuc lithologic types alternate (Billings, 1937, p. 476). The metamorphism in this belt is low grade; that is, the rocks lie in the chlorite zone.

The principal rocks are fine-grained massive to schistose, and consist of soda-rhyolite, soda-rhyolite with bluish quartz grains an eighth to a fourth inch in diameter, soda-rhyolite volcanic conglomerate, and chlorite schist. These rocks are believed to have been volcanic tuffs and breccias; flows were distinctly subordinate. Slate is locally abundant; an unusually thick bed of slate is found at the base of the Ammonoosuc volcanics a mile N. 10°W. of Woodsville on the steep slopes leading up to Gardner Mountain.

As these rocks are identical with those in the Ammonoosuc volcanics in the Littleton-Moosilauke area (Billings, 1937, p. 475–478), detailed petrographic descriptions are not given in the present paper.

The Ammonoosuc volcanics are estimated to be about 2000 feet thick in the Littleton-Moosilauke area(Billings, 1937, p. 480). The belt of Ammonoosuc volcanics northeast of Woodsville is continuous with the type locality in the Littleton quadrangle (Fig. 2). The Ammonoosuc is pre-Silurian and in recent years has been tentatively assigned to the Upper Ordovician (Billings, 1937, p. 480).

Partridge Formation

The Partridge formation, covering about 0.8 square mile, 2.5 miles northeast of Woodsville, is chiefly black slate, although locally there are thin beds of fine-grained, light-colored quartzite. Inasmuch as the top of the formation is not exposed in the Woodsville quadrangle, the thickness cannot be calculated. Billings (1937, p. 481) estimates the maximum thickness to be 2000 feet, but in many places the formation is absent due to an unconformity at the base of the Silurian formations (Billings, 1937, p. 517–518). These rocks are continuous with rocks in the Moosilauke quadrangle that have been correlated with the Partridge formation of the type locality in the Littleton quadrangle. The formation is pre-Silurian and in recent years has been tentatively assigned to the Upper Ordovician (Billings, 1937, p. 481). The age is discussed more fully later.

Clough Quartzite and Fitch Formation

The Silurian Clough and Fitch formations are not present in the Woodsville quadrangle, but they appear in adjacent quadrangles (Fig. 2). The Clough formation, chiefly quartzite and quartz conglomerate, reaches a maximum thickness of 1200 feet in the Mt. Cube quadrangle (Fig. 2), but thins northward to disappear in the Littleton quadrangle; it is Early or Middle Silurian. The Fitch formation, chiefly limestone, calcareous slate, and dolomitic slate, or the metamorphosed equivalents, is 400 to 700 feet thick; fossils indicate that it is Middle Silurian. In the Littleton-Moosilauke area a major unconformity separates the Silurian rocks from the underlying formations (Billings, 1937, p. 517–518).

Littleton Formation

In the Woodsville quadrangle the Littleton formation is found only in the extreme southeast corner, where it is chiefly gray mica schist, mica-quartz schist, and micaceous quartzite; biotite and staurolite porphyroblasts are common, although in many outcrops they have been chloritized. Detailed descriptions are given elsewhere (Billings, 1937, p. 490–491; Hadley, 1942, p. 133–134). The micaceous quartzite was formerly quarried for whetstone that was fabricated at Pike.

That portion of the Littleton formation exposed in the Woodsville quadrangle cannot be more than 2000 feet thick, but the lower part of the formation, exposed between these rocks and the underlying Fitch formation in the southwest corner of the Moosilauke quadrangle (Fig. 2), is approximately 1500 feet thick. The total exposed thickness in this area would thus be 3500 feet. In adjacent areas, the formation is 5000 feet thick.

The schists in the southeast corner of the Woodsville quadrangle are continuous with rocks in the southwest corner of the Moosilauke quadrangle that have been assigned to the Littleton formation because of their stratigraphic position (Fig. 2). Northeast of the Ammonoosuc thrust in the Littleton quadrangle, the forma-

tion has been dated by fossils as Lower Devonian (Billings, 1937, p. 494–495).

Age of Formations

The relative age of all the rocks on opposite sides of the Monroe fault cannot be determined in the Woodsville quadrangle. However, since the mapping of the Woodsville quadrangle was initiated 16 years ago, significant data have been obtained elsewhere. In the Bellows Falls quadrangle (Fig. 1), Kruger (1946, p. 169) mapped what is here called the Gile Mountain formation as the lower part of the Orfordville formation. The Meetinghouse formation is not present. He concluded that the Waits River formation was overlain in orderly succession by the Gile Mountain, Orfordville, Albee, Ammonoosuc, Partridge, Clough, Fitch, and Littleton formations (Kruger, 1946, p. 175–178). Inasmuch as the Waits River is Middle Ordovician and the Clough is Lower or Middle Silurian, he considered the intervening formations to Middle or Upper Ordovician.

Professor John Lyons of Dartmouth College has recently mapped the geology of the Hanover quadrangle. Like Kruger, he has concluded that the Waits River formation is overlain in succession by the Gile Mountain, Meetinghouse, Orfordville, Albee, Ammonoosuc, and Patridge formations. Moreover, he concludes that all these formations cannot be older than Middle Ordovician and are pre-Silurian.

The conclusions of Doll (1943a; 1943b) that the Waits River formation is Silurian and/or Devonian and that the Gile Mountain and Meetinghouse formations are Devonian are here rejected. This problem will be discussed in a later publication.

Plutonic and Hypabyssal Rocks

General Statement

The plutonic and hypabyssal rocks of the Woodsville quadrangle have been assigned to two magma series. The following, placed as nearly as possible in chronological order, are assigned to the Late Devonian (?) New Hampshire magma series: (1) metamorphosed mafic dikes, including the Leighton Hill dike complex; (2) leucocratic dikes; (3) Bethlehem gneiss; (4) French Pond granite; (5) Haverhill granodiorite; and (6) Ryegate granodiorite. A few camptonite dikes are assigned to the Mississippian (?) White Mountain magma series.

Metamorphosed Mafic Dikes and Leighton Hill Dike Complex

Metamorphosed mafic dikes are present throughout the Woodsville quadrangle. East of the Ammonoosuc thrust, they nowhere constitute more than a few per cent of the bedrock. West of this thrust, they make up 15 per cent or more of the bedrock in much of the area underlain by the Albee formation. Areas where they are even more abundant, making up 20 to 50 per cent of the bedrock, are shown on Plate 1 by a special overprint. These areas are located as follows: (1) a belt 2 miles long north of the village of Haverhill; (2) a belt 3 miles long north of the village of Newbury; and (3) a belt 13 miles long that extends from the north edge of the quadrangle along the east side of the Monroe fault as far south as Leight on Hill, whence it trends southeasterly to beyond Town Farm Hill.

In an area of extraordinary dike intrusion extending south from Boltonville, distinguished as the Leighton Hill dike complex (Pl. 1), all the outcrops are metamorphosed mafic rocks, and no exposures of metasediments were encountered. Outcrops a few tens of feet apart generally contain rocks of different structure and texture. Locally contacts between different types may be observed. For this reason the rocks in this area are believed to represent a dike complex rather than a single massive body of metamorphosed igneous rock. The eastern and southern boundaries of the area as mapped are gradational, but the western boundary is abrupt.

The metamorphosed mafic dikes are uncommon west of the Monroe fault. A few were observed in the eastern part of the Gile Mountain formation, but farther west they are rare.

The dikes range in width from a fraction of an inch up to as much as 50 feet. Most of them are parallel to the schistosity of the adjacent country rock. They characteristically cross-cut the bedding. Although a few appear to be folded, many of them truncate the fold axes,

or lie parallel to the axial planes of folds (Fig. 3).

The mafic dikes include a wide diversity of rock types, ranging from chlorite schist through actinolitic greenstones to amphibolite; textures are massive to schistose, and colors range from gray or dark-green to black. Billings and White (1950) concluded that all these rocks were originally of basaltic composition, and that the present variety of rock types is the result of differences in the thermal metamorphism and of changes in chemical composition during metamorphism, chiefly variations in the amount of introduced carbon dioxide.

Leucocratic Dikes

About 50 schistose to massive, gray or brown micaceous dikes were seen in the Gile Mountain formation. Most of these dikes are about 4 feet wide, but locally they are over 20 feet wide. They are essentially parallel to the bedding and may be sills. One appears to be folded. Essential minerals are sodic oligoclase, quartz, and muscovite, with subordinate biotite and carbonate. Inasmuch as some of these dikes are schistose, they are either pre-tectonic or syntectonic.

Bethlehem Gneiss

The Bethlehem gneiss covers approximately 0.5 square mile along the east side of the quadrangle a mile north of Pike. This is the southwestern extremity of a body 11 miles long and 1 to 2 miles wide that extends northeasterly into the Moosilauke quadrangle (Fig. 2). The Bethlehem gneiss is a gray, foliated, granulated quartz monzonite or granodiorite, composed on the average of about 30 per cent quartz, 30 per cent oligoclase-andesine, 20 per cent potash feldspar, 17 per cent biotite, 3 per cent muscovite, and traces of apatite, sphene, and pyrite. More complete descriptions and chemical analyses may be found elsewhere (Billings, 1937, p. 504–506).

Between Pike and the southern border of the quadrangle, the Littleton formation is intruded by several sills 1 to 5 feet thick. These sills are composed of a weakly foliated, fine-grained, granulated gray rock that contains biotite flakes a fraction of a millimeter in diameter. Microscopic study shows that the texture is granoblastic, with a grain size of about 0.1 mm. The mode of the one thin section studied is: oligoclase (An_{30}), 50 per cent; quartz, 39 per cent, biotite, 8 per cent: chlorite, 2 per cent; magnetite, 1 per cent; apatite, tr; allanite, tr. Hadley has described similar rocks immediately to the south in the Mt. Cube quadrangle and considers them to be an early phase of the Bethlehem gneiss (Hadley, 1942, p. 146–147). He has also shown that they cut the amphibolite sills (Hadley, 1942, Fig. 4).

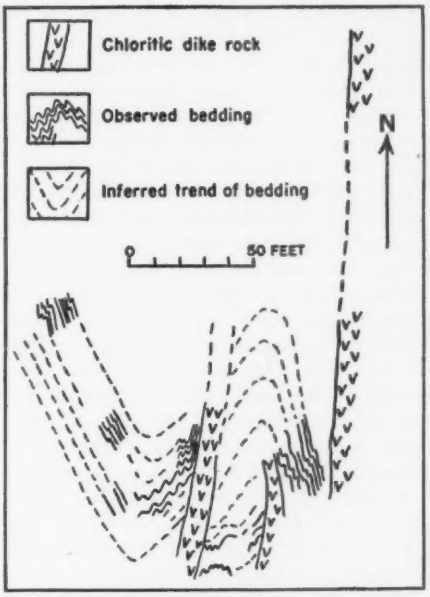


FIGURE 3.—MAFIC DIKES PARALLEL TO THE AXIAL PLANES OF FOLDS
Outcrops on slope 2000 feet south-southeast of Harriman Pond. Minor folds plunge 75°S. 10°E.

French Pond Granite

The bedrock exposed over an area of about 3 square miles to the north and east of North Haverhill is French Pond granite. Similar rocks extend eastward for 2 miles into the Moosilauke quadrangle (Fig. 2). To the south of the exposures of granite in both the Woodsville and Moosilauke quadrangles, there is a large area, 2.5 miles east-west and 1 to 1.5 miles north-

south, completely covered by glacial drift. The distribution of glacial eratics, the structure of the adjacent schists, the distribution of small dikes and sills of French Pond granite in the surrounding schists, and the intensity of the metamorphism give clues, however, to the approximate location of the southern boundary of the granite. A map of the whole stock has been published elsewhere (Billings, 1945, p. 58).

The French Pond granite is a heterogeneous unfoliated rock. The most extensive variety in the Woodsville quadrangle is a porphyritic to coarse-grained biotite granite. The phenocrysts, composed of potash feldspar, are 1 or more centimeters long; locally they are so abundant that the groundmass disappears and the rock becomes a coarse-grained granite. A second and less common variety is pink or gray, medium-grained to fine-grained biotite granite. These two varieties of the French Pond granite, separately designated on Plate 1 have been fully described elsewhere (Billings, 1937, p. 509). A third variety, found in only a few outcrops 0.5 mile due north of B.M. 497, north of North Haverhill, is a medium-grained biotite-quartz syenite with the following average mode; potash feldspar, 68%; biotite, 18%; milky quartz, 10%; sphene, 2%; and epidote, 2%.

Haverhill Granodiorite

The Haverhill granodiorite lies in the southern part of the quadrangle in the vicinity of Black Hill. The main body is oval in plan, with a north-northeast axis 1.2 miles long, and a minor axis 0.4 mile long. Small sills, however, are found to the west and south. Exposures are very good on the west, north, and east slopes of Black Hill, but elsewhere they are poor. There are two abandoned quarries within this body, one on the east slope of Black Hill and the other on the northwest slope.

The Haverhill granodiorite is typically a white granular rock composed of white feldspar, milky quartz, biotite, and muscovite. Although individual grains do not exceed 3 mm. in diameter, even a casual inspection with the hand lens shows that the texture is granoblastic and that the original grains may have been as much as 10 mm. in diameter. Generally massive, the rock locally possesses a weak foliation that strikes N. 10° E. and dips 70° W.

A much finer-grained and grayer facies is exposed on the eastern margin of the body and in sills to the west and south of the main body. In this finer facies the individual grains are

TABLE 7.—ESTIMATED MODES OF HAVERHILL GRANODIORITE

	1	2	3	4	5
Quartz	30	31	26	35	31
Plagioclase	42	43	52	45	45
Microcline	15	15	5	7	10
Biotite	2	4	10	5	5
Muscovite	3	3		5	3
Epidote	3	3	3		2
Sphene	1	1	1		1
Chlorite	4			3	2
Zircon			tr		tr
Magnetite			3	tr	1
Anorthite content of plagioclase	30	30	35	27	30

tr—present but less than 1/2 per cent

1. Main phase, altitude of 840 feet on west slope of Black Hill.
2. Main phase, altitude of 990 feet on west slope of Black Hill.
3. Contact phase, quarry at altitude of 1180 feet on east slope of Black Hill.
4. Sill in Piermont member of Albee formation, altitude of 800 feet on west slope of Black Hill.
5. Average

0.25 mm. to 1.0 mm. in diameter. This phase, particularly well exposed in an abandoned quarry at an altitude of 1180 feet on the east slope of Black Hill, appears to grade westward into the more typical main phase.

Four representative modes are given in Table 7. All four individual specimens fall within the classification of granodiorite. Microscopic study shows that the texture is essentially granoblastic, although locally there is a suggestion of hypidiomorphic granular texture. The quartz grains commonly show strain shadows. The texture thus indicates that the rock has been subjected to post-consolidation deformation, accompanied by granulaton and recrystallization.

Hadley has described Haverhill granodiorite from two small intrusive bodies in the Mt. Cube quadrangle (Hadley, 1944, p. 148).

TABLE 8.—MODES OF THE RYEGATE GRANODIORITE AND ASSOCIATED ROCKS
(Rosiwal analyses, volume per cent)

	1	2	3	4	5	6	7	8	9	10	11	12
Number of sections	5	1	1	1	7	9	2	5	4	3	1	1
Quartz	26	22	24	16	27	25	30	25	27	36	22	14
Plagioclase	51	52	64	47	61	54	36	58	45	35	44	51
Microcline	15	12	9	28	2	13	24	x	14	24	20	23
Biotite	6	12	2	6	9	7	3	10	7	3	12	9
Muscovite	2	2	1	3	1	1	7	6	7	2	2	3
Chlorite	x		x		t	x	t	x	t	t		
Allanite	x	x	x	x	x	x		x	t	x		
Epidote	x	x	x	x	x	x	t	t	t	x		
Sphene	x	x	x	x	x	x	t	x	x	x	x	x
Apatite	x	x	x	x	x	x	x	x	x	t	x	x
Zircon	t			x		t			t	t	x	
Carbonate						t		1	t			
Metallic opaque						t	t	x	t	x	x	
Per cent of anorthite in plagioclase	20	26	24	22	27	26	16	28	23	13	15	30

Grain size: matrix 0.1–0.5 mm.; phenocrysts of quartz and plagioclase, 1.0–5.0 mm.; microcline in porphyry, phenocrysts up to 15.0 mm.

x—present in 50 per cent or more of thin sections.

t—present in less than 50 per cent of thin sections.

Blank—not seen in any of thin sections.

1. Granodiorite, Blue Mountain
2. Fine-grained foliated granodiorite, Blue Mountain
3–4. Porphyritic granodiorite, Blue Mountain
5. Tonalite (quartz diorite), Groton area
6. Granodiorite, Groton area
7. Leucocratic quartz monzonite, Groton area
8. Tonalite dikes in Waits River formation
9. Granodiorite dikes in Waits River formation
10. Leucocratic quartz monzonite dikes in Waits River formation
11. Barre "granite", Rock of Ages quarry (Maynard, 1934, p. 147)
12. Barre "granite", Pirie quarry (Shimer, 1943, p. 1062)

Ryegate Granodiorite

Granodiorite is the most abundant rock in the complex of small plutons here called the Ryegate granodiorite. Associated rocks range in composition from quartz diorite to quartz monzonite.

These plutonic rocks are typically hypidiomorphic granular aggregates of oligoclase (near andesine) and quartz, with subordinate biotite and microcline. A small amount of the granodiorite is porphyritic, with microcline phenocrysts. Muscovite is present in small amounts and sphene, apatite, and zoned allanite are typical accessories. Tourmaline was not seen in any thin-sections. Epidote and chlorite are common secondary alteration minerals. Modes of the rocks, calculated on a Rosiwal stage, are presented in Table 8.

The plagioclase occurs as idiomorphic grains 1 to 5 mm. across, and quartz as allotriomorphic grains almost as large. A finer-grained equigranular aggregate of the same minerals surrounds the larger crystals and contains the biotite, muscovite, and accessories. The larger plagioclase crystals are zoned, and commonly the core contains 10 to 15 per cent more anorthite than the outer zone. Microcline is interstitial to all other minerals, except in the porphyritic rocks. In these it forms large idiomorphic phenocrysts up to 15 mm. in size. The typical interstitial character of the microcline

suggests that it is always a late mineral, and it is possible that the phenocrysts are formed by replacement at a late magmatic stage. Plagioclase in contact with microcline is myrmekitic.

Some of the rocks are foliated, and in these

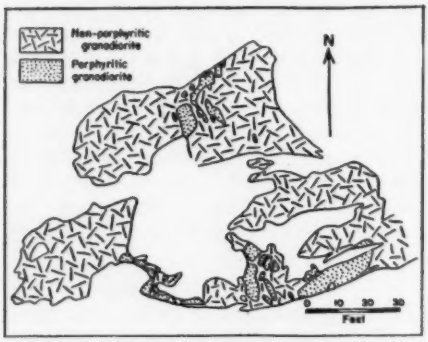


FIGURE 4.—INCLUSIONS OF PORPHYRITIC GRANODIORITE IN NON-PORPHYRITIC GRANODIORITE

Blank areas are glacial drift. Outcrop at altitude 1780 feet on the north ridge of Blue Mountain.

the degree of orientation of the biotite determines the perfection of the foliation. This planar element is generally best developed near contacts and is parallel to them, even where contacts cut across the foliation of the country rock. There is no evidence of cataclastic texture or augen structure, and the foliation is therefore interpreted as a primary flow structure.

The Ryegate granodiorite forms small plutons in two general areas, distinguished here as the Blue Mountain area and the Groton area. Blue Mountain is underlain by several subconcordant bodies of non-porphyritic granodiorite with subordinate amounts of porphyritic granodiorite. The two types have been separated in Plate 1. The porphyritic rock is found as distinct inclusions in the other (Fig. 4), and is therefore older.

A very poorly exposed complex of calcareous schists and plutonic rocks underlies the large drift-covered area, here called the Groton area, west of South Ryegate (Pl. 1). Outcrops are too few to justify drawing continuous boundaries between plutonic and metamorphic rocks. The greater part of this area is believed to be underlain by massive granodiorite.

The plutonic rocks of the Groton area include several distinct types, but little is known of their areal distribution because of the paucity of outcrops. The oldest intrusive rocks are coarse, slightly foliated granodiorite and quartz diorite. The commonest rock is a massive coarse-grained to medium-grained granodiorite

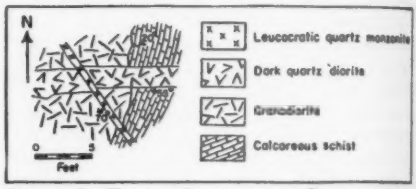


FIGURE 5.—TYPICAL RELATIONS OF GRANODIORITE AND ASSOCIATED DIKES IN THE GROTON AREA

Outcrop in the bed of the South Branch of the Wells River at altitude 1000 feet. Bedding dips 20°ESE. Arrows with numbers give direction and amount of dip of dikes.

or quartz diorite, identical in appearance with the nonporphyritic rocks of Blue Mountain. Locally the massive rock cuts the foliated, but the two types may well be penecontemporaneous. In both types, all gradations are found from quartz diorite to a granodiorite near quartz monzonite in composition.

The coarse-textured rocks are cut by dikes, up to 5 feet wide, of dark fine-grained quartz diorite, and these in turn are cut by very light-colored quartz monzonite dikes (Fig. 5).

Dikes closely related to these plutonic rocks in mineral composition, texture, and structural relations are found throughout the Waits River and Gile Mountain formations in the Woodsville quadrangle. Most of them are less than 10 feet wide, but one body about 500 feet wide was found a mile east of Ricker Mills. Areas in which these dikes are particularly abundant are distinguished by an overprint in Plate 1.

The close similarity between the Ryegate granodiorite of the Woodsville quadrangle and the Barre "granites," exposed in the northwest part of the East Barre quadrangle (Fig. 1), is suggested by the comparative modal analyses presented in Table 8. Spectrographic analysis of minor elements by Shimer (1943, p. 1055, 1057, 1060) shows an even closer correspondence between rocks from the two areas. That the two are comagmatic seems beyond serious question, inasmuch as both immediately followed the

principal orogeny of the region (Balk, 1927, p. 68).

Camptonite

Nine post-metamorphism lamprophyre dikes were found in the area. All are less than 3 feet across. Six were intruded into granodiorite. All but two strike within 8° of N. 68°E., and all dip within 20° of vertical.

Four are camptonites, in which anhedral crystals and aggregates of augite occur in a finer-grained matrix of idiomorphic grains of brown hornblende and plagioclase. Magnesite (possibly dolomite) and serpentine occur as pseudomorphic aggregates after olivine, and as scattered irregular interstitial material throughout the rock

Camptonite dikes in the Littleton-Moosilauke area were considered by Billings (1937, p. 517) to be related to the White Mountain magma series; the dikes of the Woodsville area are undoubtedly the same.

Relative Ages of Plutonic and Hypabyssal Rocks

The arrangement of the plutonic and hypabyssal rocks in chronological order is based primarily on their relation to metamorphism and the extent to which they have been deformed. Only rarely do the rocks belonging to different map units come into contact. The metamorphosed mafic dikes are considered to be oldest because they are older than the metamorphism. The leucocratic dikes are also among the oldest because they are schistose. Whether they are older or younger than the metamorphosed mafic dikes is uncertain. The Bethlehem gneiss is younger than the metamorphosed mafic dikes (Hadley, 1942, Fig. 4, p. 147) and is considered to be syntectonic (Billings, 1937, p. 536–538). The French Pond, Haverhill, and Ryegate are late tectonic or post-tectonic, as they are relatively undeformed, but their age relative to each other is uncertain.

Structure

General Statement

Inasmuch as plutonic rocks are distinctly subordinate to stratified rocks in the Woodsville quadrangle, the structural pattern is dominated by folds and faults. Three master faults—the Northey Hill, Ammonoosuc, and Monroe—trend north-northeasterly and divide the area into four major tectonic units (Fig. 2; Pl. 1).

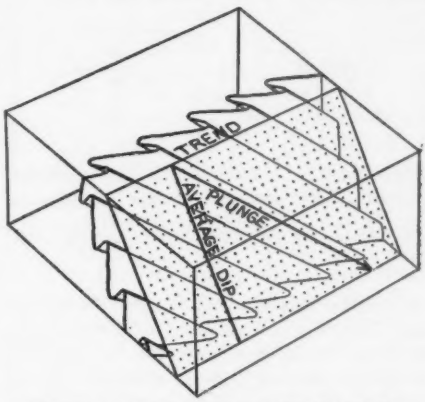


FIGURE 6.—STEREOGRAM OF FOLDED BEDDING
Sketch illustrates the use of terms "trend" and "average dip" in the description of folded bedding.

Because these three faults are so significant, they will be described first. The structure within each of the three easterly tectonic blocks is much more easily understood if a larger region is studied and for this reason Figure 2 has been prepared. Although minor folds and cleavage are present throughout the quadrangle, they were extensively studied only west of the Connecticut River.

Terminology

Trend and average dip.—In highly folded rocks the bedding strikes in various directions in different parts of an outcrop, and the only significant measurement that can be made is on the "trend" of the bedding (Balk, 1936, p. 702). The trend of bedding, as used here, is the strike of a plane that is tangent to the noses of the minor anticlines shown by a single bed. The dip of this plane is here called the "average dip" (Fig. 6). Many of the symbols for the attitude of bedding in Plate 3 actually represent trend and average dip, where no uniform strike could be measured.

Dextral and sinistral fold patterns.—The "pattern" of minor plunging folds refers to the zig

zag trace of a folded bed in an exposure (Fig. 7). Where opposite limbs of folds are of unequal length, two types of fold pattern are possible, and each type is the mirror image of the other. It is proposed to distinguish the two by the terms "dextral" and "sinistral" (Fig. 7). In

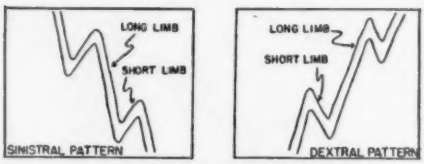


FIGURE 7.—PATTERNS OF FOLDED BEDDING IN PLAN

Illustrates use of "dextral" and "sinistral" patterns.

asymmetric folds with long and short limbs, the short limb may be regarded, for the purpose of definition, as an offset in the general course of bedding on the long limbs. If, as the bedding is followed away from a given point, the bedding is offset to the right, the fold pattern is dextral; if the offset is to the left, the pattern is sinistral. This terminology was suggested by Anderson's (1942, p. 55) proposed usage for transcurrent faults. If the terms should be used for fold patterns as viewed in cross-section, the orientation of the section and the point of view must be stated. The pattern in plan is meant here unless a section is specified.

Cleavage.—"Cleavage" is used throughout this paper to refer to the general capacity of rocks to part along essentially parallel surfaces of secondary origin. It does not include parallel joints of the order of a foot or more apart, nor bedding fissility, nor the primary flow structure of igneous rocks, which are included in the broader terms "foliation," or "foliate structures." Two prominent types of cleavage, schistosity and slip cleavage, are recognized in the Woodsville quadrangle; "cleavage" is used to refer to both collectively.

Schistosity.—"Schistosity" here means the "tendency to a common parallel orientation of the crystal elements," an arrangement that "imparts to the rock a schistose or fissile property" (Harker, 1939, p. 194). The rocks here called "schists" commonly display also a lamination of more and less micaceous layers that may generally be of tectonic origin (see "gneissic structure" of Swanson, 1941, p. 1447); but, as used here, "schistosity" refers primarily to the planar parallelism of individual mineral grains.

Slip cleavage.—"Slip cleavage" (Dale, 1899, p. 209) is a direction of splitting, transverse to schistosity, that is due to the tiny crinkles or microfolds on the schistosity surfaces of foliate rocks. Rocks may cleave along the limbs of these microfolds, on surfaces parallel to the axial planes, whether or not the limbs are actually faulted; but in its most characteristic development, slip cleavage follows the tiny faults or foliation planes that form where the limbs of microfolds are sheared off. The nature of this cleavage in east-central Vermont, and the reasons for using "slip cleavage" instead of "fracture cleavage" for the phenomenon, have been discussed elsewhere (White, 1949).

Major Faults

Northey Hill thrust.—A large fault extends north-south through the village of Pike. The schists in the southeast corner of the quadrangle belong to the Littleton formation, because they overlie the Clough and Fitch formations exposed in the southwest corner of the Moosilauke quadrangle (Fig. 2). A little more than a mile west of the east edge of the Woodsville quadrangle the Albee formation is exposed. East of the Albee formation are rocks that have been assigned to the Orfordville formation. A large fault, with a stratigraphic throw of 12,000 feet, must lie between the Orfordville and the Littleton formations. It is not easy to locate this fault, however, because the schists of the Orfordville and Littleton formations are lithologically similar. The fault is placed between black schists on the west, characteristic of the Orfordville formation, and gray schists to the east, characteristic of the Littleton formation.

About 0.7 mile north-northwest of Pike, the fault coincides with the northwest contact of the Bethlehem gneiss. The northwest contact of this intrusive body may thence be traced northeasterly for 11 miles (Fig. 2). Here, as a direct continuation of the northwest contact of the Bethlehem gneiss, a large fault emerges and may be traced northward to its type locality at Northey Hill in the Moosilauke quadrangle

(Billings, 1937, p. 530–531). The large fault in the southeast corner of the Woodsville quadrangle is therefore considered to be the Northey Hill fault.

The relatively straight course of the Northey Hill fault in the Woodsville quadrangle, despite a relief of over 500 feet, implies that it is essentially vertical. Where observed at one locality in the Moosilauke quadrangle the dip is 75° NW. (Billings, 1937, p. 530). The fault is considered to be older than the metamorphism, because the metamorphism is of equal grade on both sides. Moreover, in the Moosilauke quadrangle, staurolite porphyroblasts within a few feet of the fault are undeformed. Brecciation, mylonitization, and silicification have not been observed along the fault in the Woodsville quadrangle.

Ammonoosuc thrust.—The Ammonoosuc thrust may be traced southwesterly from a point on the east side of the quadrangle 3 miles northeast of Woodsville to the south edge of the quadrangle 2 miles south of the village of Haverhill. Throughout most of its extent the fault is on the New Hampshire side of the Connecticut River, but in two places, one 1.5 miles south of Woodsville and the other 1.5 miles west of the village of North Haverhill, it is on the Vermont side of the river.

The fault was first recognized in the Littleton and Moosilauke quadrangles because of a sharp stratigraphic discontinuity where strata high in the Littleton formation are in contact with strata well down in the Ammonoosuc volcanics. There the stratigraphic separation is of the order of 7000 feet. A great deal of mechanical evidence for the fault was also discovered there; not only was the fault plane observed in several localities, but locally there is considerable mylonitization and silicification. The silicified zones were especially helpful in tracing the fault. Moreover, a sharp discontinuity in the grade of metamorphism characterizes the trace of the fault.

In the Woodsville quadrangle, stratigraphic evidence for the fault is found only in the area 2.5 miles northeast of Woodsville, where the top of the Ammonoosuc volcanics is in contact with rocks well down in the Albee formation. South of Woodsville, the Albee formation lies on both sides of the fault and stratigraphic evidence is lacking. It is therefore necessary to use the other criteria to trace the fault.

Especially pertinent data are found along the valley of Burton Brook 1.5 miles northeast of Woodsville. The valley of Burton Brook has been excavated to a depth of 100 to 200 feet for a mile along the trace of the fault. A third of a mile north of Lake Gardner, the footwall of the fault has been silicified for 2000 feet along the strike. The thickness of the silicified zone ranges from 6 to 30 feet. The average strike of the top of this silicified zone is N. 30°E. The dip of the top of the silicified zone, presumed to be the fault plane, was observed at four different localities to be 32°, 36°, 38° and 40° toward the northwest; the average is 36° NW. Near the southwest end of the silicified zone, near the road, the top of the silicified zone is a wavy surface; the wavelength of these gentle rolls is 2 to 3 feet, and the amplitude is half a foot. These rolls, plunging directly down the dip of the surface, indicate that the displacement along the fault was essentially a dip-slip movement with no strike-slip component. The evidence that the hanging wall has moved up relative to the footwall has been discussed elsewhere (Billings, 1937, p. 529).

For 3.5 miles south of Woodsville, the trace of the fault can be located only approximately, chiefly because of glacial deposits and alluvium along the Connecticut River. On the Vermont side of the river, for 1.5 miles south of Horse Meadows, the trace of the fault has been accurately located. Although the rocks on both sides of the fault belong to the Albee formation, the rocks on the east side of the fault are in the staurolite zone whereas those of the west side are in the chlorite zone.

From a point 1.8 miles southwest of the village of North Haverhill, the fault can be readily traced southward for 2 miles. The rocks east of the fault are in the staurolite zone; those on the west side are in the chlorite zone. Two miles southwest of North Haverhill, a large silicified zone occurs along the fault. This silicified zone, the north end of which lies 0.2 miles south of the Dartmouth College Road, strikes N.15°W., is 150 feet long, and 75 feet wide. The east side of the zone is a network of quartz veins, and the whole zone is apparently a silicified breccia.

For nearly 0.5 mile to the south, the trace of the fault lies in the bottom of a narrow valley.

For 2 miles south of the railroad north of Oliverian Brook, the trace of the fault is buried under glacial drift. The southernmost mile of the thrust in the Woodsville quadrangle is accurately located because of the contrast in metamorphism on opposite sides of the fault.

The retrograde metamorphism associated with the fault will be described below.

Monroe fault.—The rocks of the Albee and Orfordville formations are separated from those of the Meetinghouse slate and Gile Mountain formation along a fault that crosses the quadrangle from north to south. This fault has been named the Monroe fault (Eric, White, and Hadley, 1941). Although the strike of the bedding and schistosity next to the fault are essentially parallel to it and none of the outcrops of the contact suggest a structural break, there is evidence to prove that the boundary between the formations is a fault.

First, in the St. Johnsbury quadrangle (Fig. 2), Ammonoosuc volcanics on the east side of the fault are in contact with the Gile Mountain formation on the west. In the southern part of the Woodsville quadrangle and in the Mt. Cube quadrangle, the Orfordville formation on the east side is next to the Meetinghouse slate on the west. Moreover, the lower part of the Orfordville formation is missing at the contact. In most of the Woodsville quadrangle, phyllite of the Albee formation on the east is the dominant rock in contact with the Meetinghouse slate and Gile Mountain formation on the west, but locally, at Boltonville, quartzite predominates. Second, Eric has found detailed evidence of truncation of bedding in the Vermont portion of the Littleton quadrangle (Eric, White, and Hadley, 1941). Lastly, the biotite isograd is slightly offset in the vicinity of Boltonville.

The trace of the fault has been continuously mapped for 72 miles and probably extends much farther. Throughout most of the Woodsville quadrangle, the Orfordville formation has been cut out along the Monroe fault. Consequently the stratigraphic throw is here equal to the thickness of the Orfordville formation, approximately 5000 feet. The west side is upthrown, but the possibility of a reversal of movement subsequent to the metamorphism is discussed later.

Inasmuch as the strike of the bedding and the schistosity on the two sides of the fault are almost parallel to the strike of the fault, it is inferred that the dip of the fault is equally concordant with the dip of the bedding and schistosity (Pl. 2). Such parallelism is actually tectonic, because, on a regional scale, the contact is crosscutting. If the fault dips parallel to the bedding, the dip is 70°W. at the extreme north end of the area, vertical in most of the north half of the area, and 60°E. at the south end.

Structure East of Northey Hill Thrust

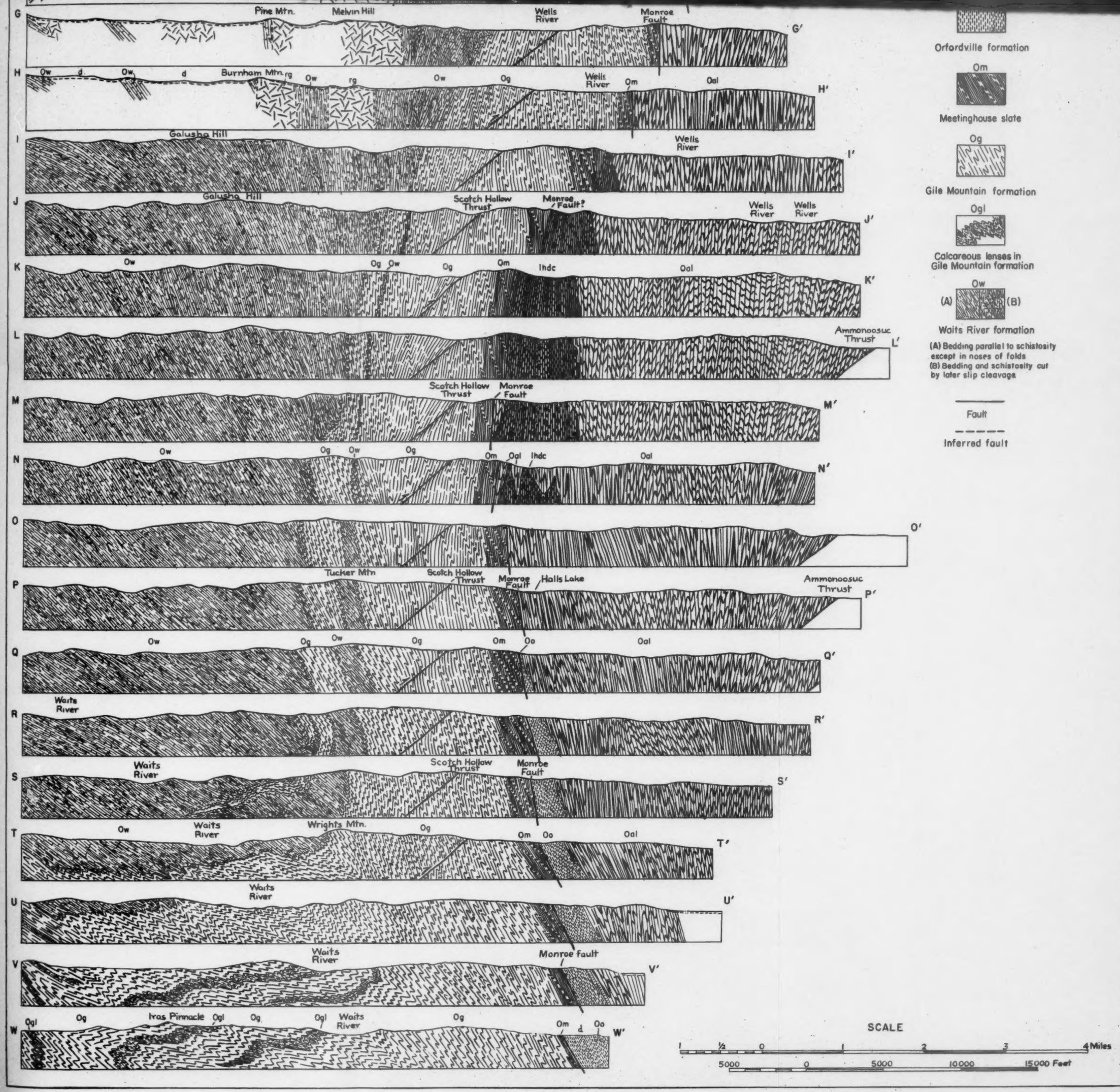
The Owls Head dome is a large dome in the southwest corner of the Moosilauke quadrangle and the northwest corner of the Rumney quadrangle (Fig. 2). The core of the dome is occupied by the Owls Head granite of the Oliverian magma series. On the west flank of the Owls Head dome, the Ammonoosuc, Clough, Fitch, and Littleton formations dip westward. The Littleton formation in the southeast corner of the Woodsville quadrangle is part of the west flank of this dome.

The Bethlehem gneiss a mile north of Pike is the southwest end of a body 11 miles long and 1 to 2 miles wide that extends northeasterly into the Moosilauke quadrangle (Fig. 2). This body, termed the Green Mountain pluton, is a concordant sheet, in the northern half of which the contacts and foliation are vertical. Toward the south, however, the contacts and foliation dip about 75°NW; in the Woodsville quadrangle the foliation and presumably the contacts dip about 60°NW. The pluton thus appears to be a huge sheet that is vertical in its northeastern half but toward the southwest gradually rolls over to dip 60°NW.

Structure between Northey Hill and Ammonoosuc Thrusts

That portion of the quadrangle east of the Ammonoosuc thrust is underlain chiefly by the Albee formation, except in the area occupied by French Pond granite. In the southern part of the quadrangle, however, the appearance of the Orfordville formation between the Albee formation and the Northey Hill fault indicates that the trough of a syncline lies somewhere to the west of the Northey Hill fault. The geological map of the Mt. Cube quadrangle (Fig. 2) (Hadley, 1942, Pl. 4)

WEST
EAST
A
A'
B
B'
Ow
rg
Ow
Og
rg
Blue Mtn.
Og
rg
Og
Og
Om
Oal
Blue Mountain
C
C'
Wells River
Ryegate anticline
Monroe Fault
D
D'
E
E'
Ow
rg
Ow
d
rg
Wells River
d
Ow
Og
Om
Oal
F
F'
Wells River
Scotch Hollow Thrust
G
G'
Pine Mtn.
Melvin Hill
Wells River
Monroe Fault
H
H'
Ow
d
Ow
d
Burnham Mtn.
rg
Ow
rg
Ow
Og
Wells River
Om
Oal
I
I'
Galusha Hill
Wells River
J
J'
Galusha Hill
Scotch Hollow Thrust
Monroe Fault?
Wells River
Wells River
K
K'
Ow
Og
Ow
Og
Om
lhdc
Oal
L
L'
Ammonoosuc Thrust
Scotch Hollow Thrust
Monroe Fault
M
M'
N
N'
Ow
Og
Ow
Og
Om
Oal
lhdc
Oal
O
O'
Tucker Mtn
Scotch Hollow Thrust
Monroe Fault
Halls Lake
Ammonoosuc Thrust
P
P'
Ow
Og
Ow
Og
Om
Oo
Oal
Q
Q'
Waits River
R
R'
EXPLANATION
d
Drift-covered, bedrock not exposed
g
Ryegate granodiorite
lhdc
Leighton Hill dike complex
Oal
Albee formation
Thin straight lines are schistosity, thick lines are bedding. Folds diagrammatic
Oo
Orfordville formation
Om
Meetinghouse slate
Og
Gile Mountain formation
Ogl
Calcareous lenses in Gile Mountain formation
Ow
(A)
(B)
Waits River formation
(A) Bedding parallel to schistosity except in noses of folds
(B) Bedding and schistosity cut by later slip cleavage
Fault
Inferred fault



SERIES OF CROSS SECTIONS OF VERMONT PORTION OF WOODSVILLE QUADRANGLE, VERMONT-NEW HAMPSHIRE

shows that the trough of the syncline lies about a mile west of the Northey Hill fault; in the southern end of the Woodsville quadrangle it would lie between Morris Brook and Catamount Ridge. The Piermont member of the Albee formation, exposed over a large area on the west limb of the syncline, pinches out eastward and does not appear on the east limb.

In the Moosilauke quadrangle this fold has been called the Salmon Hole Brook syncline.

The regional map-pattern of this fold in the Mt. Cube, Woodsville, and Moosilauke quadrangles (Fig. 2) suggests a syncline plunging toward the northeast. However, the minor folds in the Moosilauke quadrangle plunge steeply southwest, indicating an inverted plunge (Billings, 1937, p. 522). Throughout most of the Woodsville quadrangle the minor folds in this syncline plunge northwest, west and southwest; the axial planes dip steeply northwest and west. The average of all readings indicates a steep plunge to the west; thus, in general, the plunge in the Woodsville quadrangle appears to be inverted. But to the south, in the Mt. Cube quadrangle, the plunge becomes northeast and is thus normal. In the southeast corner of the Woodsville quadrangle, the Orfordville formation plunges southward beneath the Albee formation.

The Black Hill pluton, composed of Haverhill granodiorite, is a concordant lens 1.2 miles long and 0.4 mile wide that dips steeply to the west. The eastern contact, well exposed in a quarry at an altitude of 1180 feet on the east slopes of Black Hill, strikes N. 30°E. and dips 54°NW. parallel to the foliation in the rocks on both sides. The foliation throughout most of the pluton dips 75°W. This is true even near the western contact, and indicates that this contact dips at a similar angle. For 100 to 200 feet west of the main body, sills of Haverhill granodiorite 5–10 feet thick intrude the schists of the Piermont member of the Albee formation. On the ridge 1 mile south-southwest of Black Hill, there are sills of Haverhill granodiorite in the schists; this suggests that the pluton may fray out at its southwest end into a series of sills.

The French Pond granite forms a large cylindrical pluton with a vertical axis. A detailed map has been presented elsewhere (Billings, 1945, p. 57).

Structure between Ammonoosuc and Monroe Faults

Major structure.—The outcrop pattern of the formations between the Monroe fault and the Ammonoosuc thrust in the northern part of the Woodsville and southeastern St. Johnsbury quadrangle (Fig. 2) defines two major northward-plunging synclines and an intervening anticline. The Ammonoosuc volcanics, overlain by the Partridge formation, outline the nose of a northward-plunging syncline, the Walker Mountain syncline (Billings, 1937, p. 519–521), just north of the village of Woodsville. An oval-shaped area of Ammonoosuc, bounded on the west by the Monroe fault in the southeast ninth of the St. Johnsbury quadrangle, reveals a northeasterly plunging syncline called the Monroe syncline. The anticline between these synclines, the Gardner Mountain anticline, has its crest in a broad area underlain by the Albee formation. The axial planes of all these folds may be projected for a short distance into the central part of the Woodsville quadrangle, but cannot be readily traced through the area of complexly folded Albee formation that occupies the whole space between the Monroe fault and the Ammonoosuc thrust southwest of Woodsville.

Near the southern border of the quadrangle, just east of the Connecticut River, the Orfordville formation is exposed in the core of a northward-plunging anticline that Hadley (1942, p. 154) has called the Piermont anticline. The Orfordville is also exposed in a narrow belt that extends along the east side of the Monroe fault south from West Newbury. There is presumably a tight syncline, therefore, in the Albee formation between these two areas of the Orfordville.

Although the Hardy Hill quartzite is well exposed on the southeast limb of the Piermont anticline, it does not appear on the northwest limb. Locally the lower slates of the Orfordville formation—those beneath the Hardy Hill quartzite—are in contact with the Albee formation. These relations indicate that the northwest limb of the anticline is broken by a fault, which is considered to be a thrust dipping steeply to the southeast.

Correlation of the anticline and syncline at the southern border of the quadrangle with the

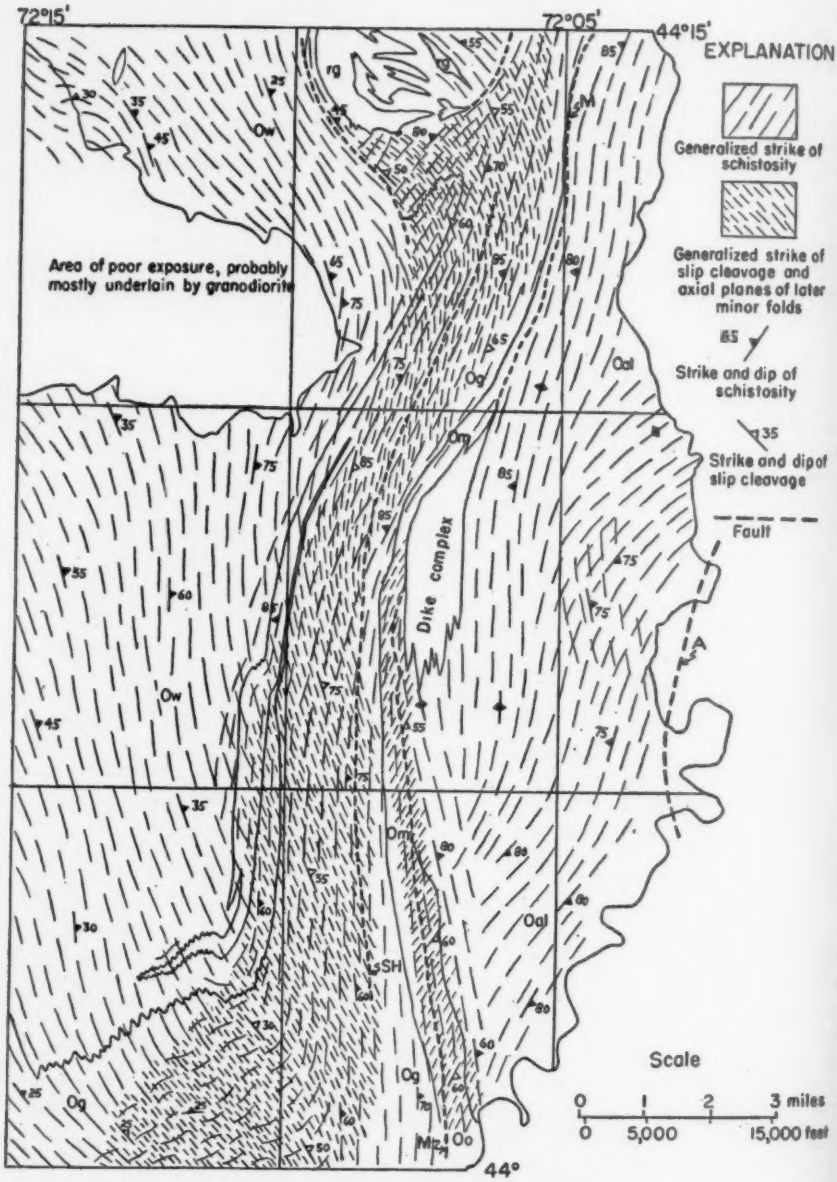


FIGURE 8.—GENERALIZED STRIKES OF CLEAVAGES

Lines based on data of Plate 3. Lithologic units and faults identified by letter symbols as follows: Ow, Waits River formation; Og, Gile Mountain formation; Om, Meetinghouse slate; Oo, Orfordville formation; Oal, Albee formation; rg, Ryegate granodiorite; SH, Scotch Hollow thrust; M, Monroe fault; A, Ammonoosuc thrust.

folds in the northern part is not established. If the axial planes of all the major folds are more or less parallel to the trend of the schistosity in the Albee formation (Fig. 8), the Walker Mountain syncline may be the same as the syncline west of the Piermont anticline. Under this interpretation, the Monroe syncline and Gardner Mountain anticline die out or are cut off towards the south by the Monroe fault, and the Piermont anticline is cut off by the Ammonoosuc thrust somewhere near Newbury. Under any alternative correlation of major folds, the schistosity in the Albee formation must be assumed to cut across the axial planes of the major folds, and would presumably, therefore, be later than the major folds.

Minor structural features.—Minor structural features of the rocks include minor folds, schistosity, slip cleavage, and shear planes. Most outcrops of the Albee formation contain minor folds that range from minute crenulations in the bedding of phyllite to folds 10 feet or more in amplitude in massive quartzite. In phyllite, the folds are nearly isoclinal and of the type called similar folds; their axial planes are essentially parallel to the schistosity in the vicinity. In massive quartzite, the folds are more open (Fig. 9) and approach parallel or concentric folds in shape; their axial planes may depart by as much as 30° from the attitude of the schistosity in the vicinity.

The axial planes of minor folds (Pl. 3) strike north to northeast, and dip vertically or steeply east; northwest dips are less common. The axes of minor folds have uniformly steep plunges, either northward or southward. Relatively few folds plunge at angles less than 60°; a large number plunge directly down the dip of their axial planes; and none were observed to have horizontal or near-horizontal axes. Both north and south plunges may be found in the same or adjacent outcrops, and at one place a single curved fold axis was seen to plunge northward at the top of a small cliff, vertically in the central part, and southward at the bottom. Inasmuch as the regional plunge is northward, as revealed by the outcrop pattern of older and younger formations in the major folds (Fig. 2; Pl. 1), and inasmuch as all evidence suggests that the minor folds reverse their direction of plunge through the vertical rather than the horizontal, the southward-plunging axes are believed to be overturned. Minor folds that appear to be southward-plunging anticlines are actually synclines in the sense that they have younger rocks in their cores than on their flanks. The word "overturned," as applied here to fold

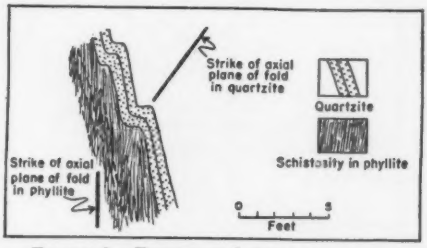


FIGURE 9.—EFFECT OF LITHOLOGY ON THE ORIENTATION OF MINOR FOLDS IN THE ALBEE FORMATION

Diagrammatic plan, folds plunge vertically. Heavy dashed lines are bedding in phyllite.

axes, is intended only as a descriptive term, because it is not at all clear whether these fold axes have actually been rotated from a subhorizontal to a steep or overturned position, or whether the folds were originally formed with essentially their present attitude, perhaps in response to differential movement parallel to the strike of the rocks.

The patterns of minor folds west of the Connecticut River are shown in Figure 10. A slight majority of those found east of the Monroe fault are sinistral.

The dominant cleavage in the rocks is schistosity. In thick phyllite beds, the schistosity is due to nearly perfect orientation of the micaceous elements, and is parallel to the axial planes of minor folds. The quartzites of the Albee formation have no cleavage, although locally there are fan joints parallel to the axes of folds and more or less normal to bedding. Where quartzite and phyllite (or schist) are thinly interbedded, the relation of schistosity to bedding differs in the various parts of folds. In the noses of folds, the schistosity is parallel to the axial plane where the phyllite beds are more than 3 or 4 inches thick; it approaches parallelism with the limbs away from the noses of folds (Fig. 11), and, in outcrops with no folds, the schistosity is essentially parallel to bedding. The micas within 2 or 3 millimeters of quartzite

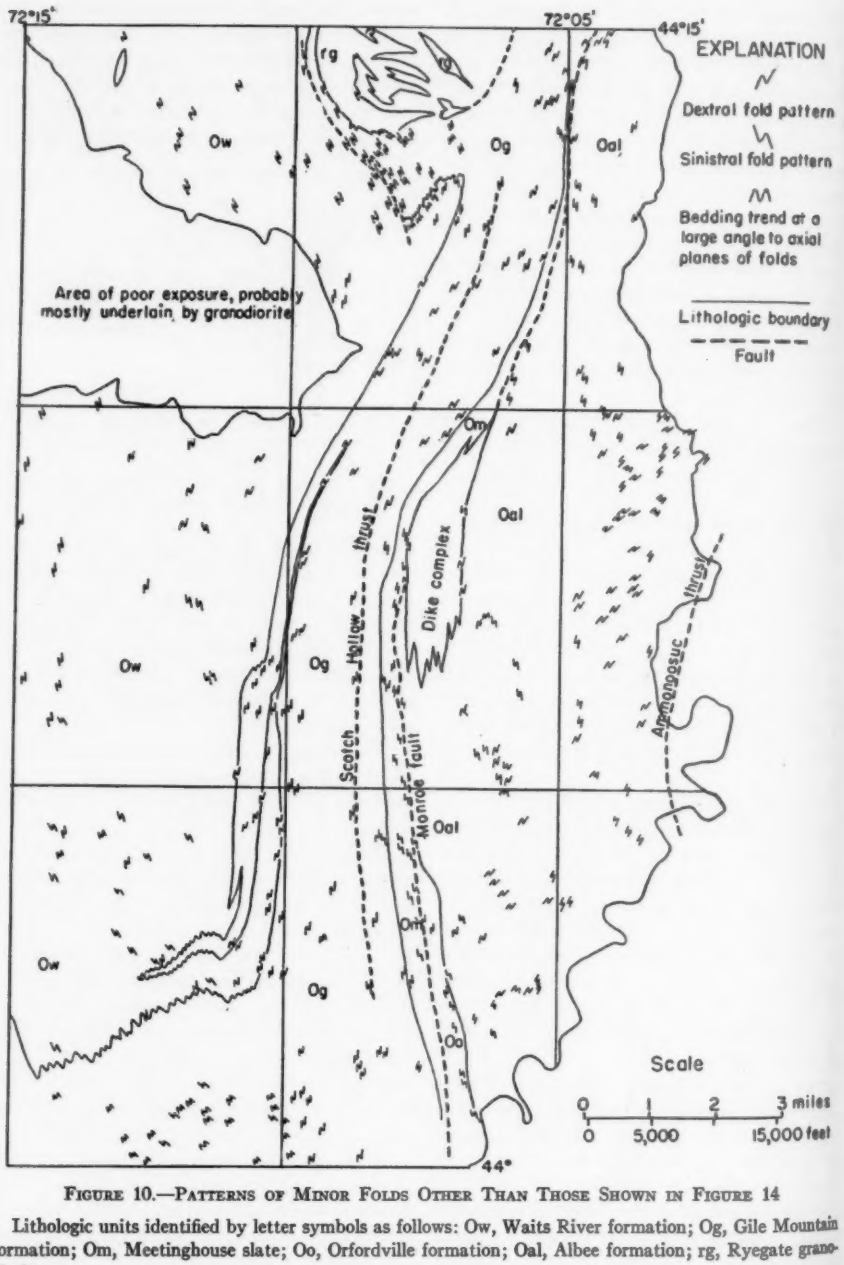


FIGURE 10.—PATTERNS OF MINOR FOLDS OTHER THAN THOSE SHOWN IN FIGURE 14

Lithologic units identified by letter symbols as follows: Ow, Waits River formation; Og, Gile Mountain formation; Om, Meetinghouse slate; Oo, Orfordville formation; Oal, Albee formation; rg, Ryegate granodiorite.

beds are parallel to bedding in all parts of folds, as are the micas in thin micaceous partings between quartzite beds.

Slip cleavage is found in rocks of both the Albee and Orfordville formations in the vicinity of the Monroe fault in the southern half of the quadrangle (Fig. 8; Pl. 3). Near the fault, schistosity and bedding strike parallel to the fault and dip steeply east. Slip cleavage, transecting both schistosity and bedding, strikes northeast and dips steeply southeast. It is parallel to the axial planes of minor folds that differ from the more common type described; in the minor folds associated with slip cleavage, both schistosity and bedding wrap around the noses. These folds plunge steeply east to southeast (Pl. 3).

Two possible explanations of the slip cleavage and associated minor folds are offered here. Both are suggested by the fact, emphasized by trend lines in Figure 8, that the schistosity near the Monroe fault in the southern part of the quadrangle is parallel to the fault and seems to lie athwart the northeastward trend of schistosity farther east, whereas the slip cleavage seems to lie parallel to the projected trend of this northeast-striking schistosity. First, the schistosity that is parallel to the fault may be slightly older than the dominant schistosity farther east, and perhaps due to the movements on the fault itself. Later movements that gave rise to the dominant northeastward-trending schistosity would then have acted, in the vicinity of the fault, on rocks already schistose, and formed slip cleavage by redeforming them.

An alternative, and perhaps less likely explanation involves the assumption that an earlier ubiquitous schistosity, oriented more or less parallel to the Monroe fault, has been destroyed by a later northeast-trending schistosity, except in the vicinity of the fault. Near the fault, the movements that produced the later schistosity must have been capable only of superimposing a slip cleavage on the earlier schistosity.

Shear planes with small to large displacement are characteristic of much of the Albee formation. They are particularly conspicuous in large outcrops of interbedded phyllite and quartzite (Fig. 12), where they are revealed by offsets in the bedding. These shear planes are generally parallel to the schistosity in strike and dip, and may be inconspicuous where both walls are phyllite. The movement on some of these shear planes must have taken place after the full development of the schistosity, because in a few places the schistosity, intersecting the shear planes at a low angle, is noticeably bent at the intersection. Movement on these shear planes may have been an important part of the total deformation of the rocks, particularly in the closing stages of the orogeny.

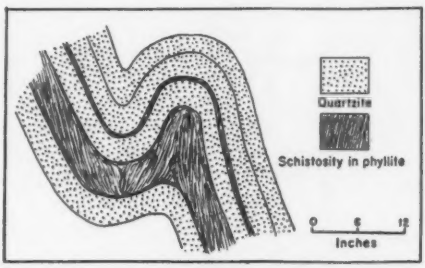


FIGURE 11.—RELATION OF SCHISTOSITY TO QUARTZITE BEDS IN MINOR FOLDS IN THE ALBEE FORMATION

Diagrammatic plan, folds plunge vertically.

Structure West of Monroe Fault

General statement.—The outcrop pattern of formation boundaries west of the Monroe fault does not, by itself, provide sufficient evidence for understanding the major structure. Clues to the nature of the major folds and to the sequence of tectonic events are derived chiefly from the attitude and relationships of minor folds and cleavage, and for this reason minor structural features are described first.

Evidence of two distinct stages in the deformation of the rocks west of the Monroe fault will be presented. The term "stage" is used here, without any implication that separate orogenies need be involved, to separate distinct tectonic events that may well have taken place at different times in the same general orogeny. The earlier stage was characterized by the widespread formation of schistosity more or less parallel to bedding and formation boundaries, and by northward-plung-

ing minor folds with a predominantly sinistral pattern; the Ryegate anticline, just southeast of Blue Mountain, was formed during this earlier stage of deformation. The later stage of deformation was less ubiquitous than the earlier, and the structural features associated with this stage show wide variation in type and intensity of development. Cleavage, if present, may be slip cleavage or schistosity; minor folds plunge northward and have predominantly dextral patterns. The Wrights Mountain fold, in the southwest ninth of the quadrangle (revealed by the change in course, from south to west-southwest, of the boundary of the Waits River formation), belongs to this stage of deformation. Superposition of features of the later stage on those of the earlier are very conspicuous over large areas and afford the principal evidence for the distinction between the stages and their sequence.

Most of the detailed structural measurements on which the following descriptions and interpretations are based are presented in Plate 3. Figure 8 gives a more generalized view of the trends of the various cleavages observed in the area, and should be studied for proper understanding of the symbols in Plate 3 and their interpretation. It will be noted in Figure 8 that the area west of the Monroe fault may be roughly divided into three parallel belts of unequal width on the basis of the character of the cleavage in each. A central belt contains slip cleavage transecting schistosity. The schistosity in the central and eastern belts is regarded as

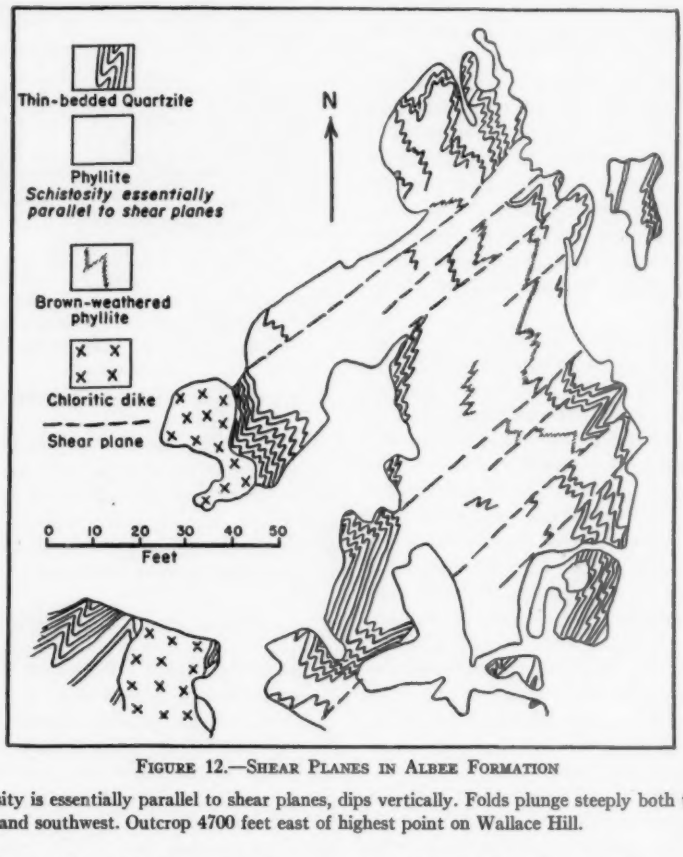


FIGURE 12.—SHEAR PLANES IN ALBEE FORMATION

Schistosity is essentially parallel to shear planes, dips vertically. Folds plunge steeply both towards the northeast and southwest. Outcrop 4700 feet east of highest point on Wallace Hill.

belonging to the earlier stage of deformation, and is called "earlier schistosity." The schistosity in the western belt, which is parallel to the projected trend of the slip cleavage, is considered to be a product of the later stage of deformation, and is called "later schistosity." Earlier schistosity was not recognized in the western belt, as will be discussed below. It seemed undesirable to distinguish the earlier and later schistosities in Plate 3 because the principal criterion for distinction is attitude; this criterion cannot be applied to all outcrops with complete objectivity in borderline cases.

Minor structural features of the earlier stage.—The characteristic cleavage of the earlier stage is a schistosity that is due primarily to a high degree of parallelism of the micaceous constituents of the rocks. To a lesser extent, it is due to dimensional orientation of tabular grains of quartz, calcite, and feldspar, where present. Schistosity is best developed in the micaceous rocks, and less conspicuous in the more quartzose or calcareous layers, but virtually all the rocks have this property in some degree. There are numerous tiny shear planes parallel to the schistosity, and most of the quartz-mica schists consist of elongate lenses or layers rich in quartz separated by thin micaceous foliae ("gneissic structure" of Swanson, 1941). But in all the rocks, quartzose or micaceous, mineral parallelism is the dominant cause of the schistosity.

There is no measurable angle between bedding and schistosity in most outcrops. Within the limits of observational errors, the average trend of the schistosity is parallel to the boundaries of formations. For practical purposes, therefore, the schistosity may be termed bedding-schistosity. As used here, the term does not imply that bedding and schistosity nowhere intersect one another, for such intersections can be seen or inferred at many places. The term simply means that, within any small area, the average attitude of schistosity is parallel to the average attitude of bedding and that the trend of bedding does not deviate appreciably from the strike of the schistosity. The general parallelism of schistosity and formation boundaries suggests that the angle between these average strikes, if any, is very small in the quadrangle as a whole.

The attitude of the earlier schistosity and bedding in individual outcrops is shown in Plate 3, and symbols for the two have been combined where reliable bedding was seen to

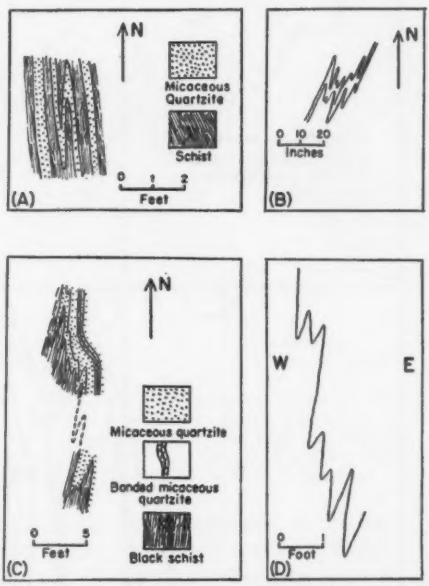


FIGURE 13.—TYPICAL EARLIER MINOR FOLDS WITH AXIAL-PLANE SCHISTOSITY

(A) Idealized earlier minor fold; axial plane steep to vertical, axis plunges 35°N. (B) Folded quartzose marble layer in schist. Outcrop in brook 4600 feet east of West Bradford School. (C) Outcrop 4700 feet northwest of Ryegate village, about 50 feet northeast of road. (D) Trace of a folded bed in a plane normal to the plunge of folds. Outcrop 2500 feet north of West Cemetery, southwest of Blue Mountain.

be parallel to schistosity. It should be re-emphasized that only the schistosity in and east of the belt of later slip cleavage (Fig. 8) is interpreted as earlier schistosity. The general parallelism of earlier schistosity and bedding and formation boundaries is evident from Plate 3, and holds even where the strike of the eastern boundary of the Waits River formation strikes east-northeast, as in the vicinity of Bradford Center.

The earlier minor folds (Fig. 13) are nearly isoclinal. The schistosity just described is parallel to their axial planes, and cuts the bedding at a high angle in their noses. On the limbs of

prominent in the more quartzose rocks, where the slip planes are more widely spaced and discontinuous (Fig. 15A), and there has been little or no bending of the pre-existing earlier schistosity planes adjacent to the slip planes.

West of the belt of slip cleavage (Fig. 8), micaceous foliae like those that define the slip cleavage farther east are more or less merged, and the micas in the intervening more quartzose lamellae or lenticles are dominantly parallel to the foliation planes. The cleavage here may, for practical purposes, be called a schistosity, although (Fig. 8) it is parallel to the projected trend of the slip cleavage and makes an appreciable angle with such earlier schistosity as can be recognized in contiguous rocks to the east. This schistosity in the western third of the quadrangle is called later schistosity to distinguish it from the earlier bedding-schistosity recognized in and east of the belt of slip cleavage. The earlier bedding-schistosity may very well have been present in this western area before the development of the later cleavage, but, if so, it was more or less completely obliterated or masked by the later.

As described in more detail elsewhere (White, 1949, p. 590–591), the evolution of schistosity as an end-product of the development of slip cleavage and the obliteration of the earlier schistosity by formation of the later, seem to have involved two distinct processes, namely rotation and smearing out of the earlier schistosity, and growth of new mica flakes oriented parallel to the later cleavage. It is generally impossible to tell, however, whether a given group of parallel mica flakes, now parallel to the later schistosity, represents mechanically reoriented earlier schistosity or new micas that have grown with an orientation controlled by the later cleavage direction.

Minor folds related to the later stage of deformation are found in most outcrops west of the Monroe fault. In and east of the belt of slip cleavage (Fig. 8), they are readily identified by the fact that the earlier schistosity wraps around the noses of the folds. They range from a few inches to about a foot across in dominantly argillaceous rocks, but may be several feet across where most of the rock is micaceous quartzite. These folds are fairly open near the Monroe fault and become tighter towards the west. Near the eastern border of the Waits River formation, the minor folds may be nearly isoclinal, although the angle between limbs is generally over 30°. In shape, the folds closely approach the type called similar folds, even in fairly massive micaceous quartzite beds (Fig. 15A). Where slip cleavage is present, it is rigidly parallel to the axial planes of the folds both in the noses and on the limbs.

West of the zone of slip cleavage, tight to isoclinal folds with axial planes parallel to the later schistosity are abundant. One or both limbs of many are sheared off. Most of these minor folds are believed to be products of the later stage of deformation, but insofar as the schistosity here identified as later schistosity may, in part, represent earlier schistosity that has been rotated, a few of these minor folds may likewise represent rotated earlier folds. As in the case of the earlier minor folds farther east, the later schistosity tends to approach parallelism with bedding on the limbs of minor folds. There is, therefore, no simple basis for distinguishing possible earlier from later minor folds in the westernmost part of the quadrangle. Hadley (1950, p. 31–32) independently arrived at an almost identical interpretation of the relation of slip cleavage and later schistosity in the Mt. Cube quadrangle.

The minor folds in calcareous rocks of the Waits River formation differ, somewhat, from those in the argillaceous rocks of the Waits River and Gile Mountain formations. In general, two types of fold may be distinguished. One type is made manifest by the contacts between dominantly calcareous and dominantly argillaceous beds. The folds brought out by these contacts are mostly 6 inches to a foot across, and range from fairly open to almost isoclinal. Where the folds are nearly isoclinal or characterized by short and long limbs, the schistosity is essentially parallel to the long limbs. At the noses of isoclinal folds and other places where contacts cut across the schistosity there is small scale "interdigitation" (Greenly, 1930, p. 177–178) of calcareous and argillaceous material.

Another type of minor fold is found only in the beds of quartzose marble, and is brought out on weathered surfaces by a faint banding that is due to slight differences in composition

of alternate layers. These folds are almost isoclinal, with amplitudes of several feet and widths of 2 to 6 inches between the same layer on opposite limbs. They may occur where the boundaries of the marble bed are perfectly straight, showing no evidence whatever of folding. The axial planes of these folds are parallel to or make a very low angle with the contacts between calcareous and argillaceous beds. Scattered mica flakes in the quartzose marble are fairly well oriented parallel to the axial planes of these isoclinal folds on the limbs and in the noses. This highly attenuated second type of later minor fold is taken as evidence that during deformation the calcareous layers were more plastic than the argillaceous layers.

It is difficult or impossible at a great many outcrops in the Waits River formation to determine the prevailing pattern of the minor folds, although their axes may be very obvious. The limbs of folds are sheared off, and only isolated noses of folds are preserved. Individual calcareous layers cannot be traced across such an outcrop. The prominent banding in such outcrops is large-scale cleavage banding rather than bedding. The preservation of noses of folds in separate cleavage bands is suggested on a highly exaggerated scale and in greatly oversimplified form in the geologic cross sections (Pl. 2).

Some of the folds in calcareous rocks are indescribably complex. A fairly common type is represented by Figure 16A; it is generally not clear in such exposures whether the tapering lenses of quartzose marble are original stratigraphic lenses, disrupted beds, or earlier folds whose axial planes have been deformed during the later stage of deformation. Even more complicated folds that defy analysis (Fig. 16B) are found in many places.

Minor folds of the later stage throughout the area west of the Monroe fault have axial planes that strike north to northwest and dip east to northeast, parallel to the later cleavage. The folds plunge uniformly northward at angles that range, for the most part, from 5° to 55°N. In the area of slip cleavage, the later minor folds are almost exclusively dextral in plan (Fig. 10). This dextral pattern suggests that during the later stage of deformation, rocks on the west moved relatively upward and northward with respect to rocks on the east, a complete reversal of the direction of differential movement indicated by the sinistral earlier minor folds. West of the area of slip cleavage,

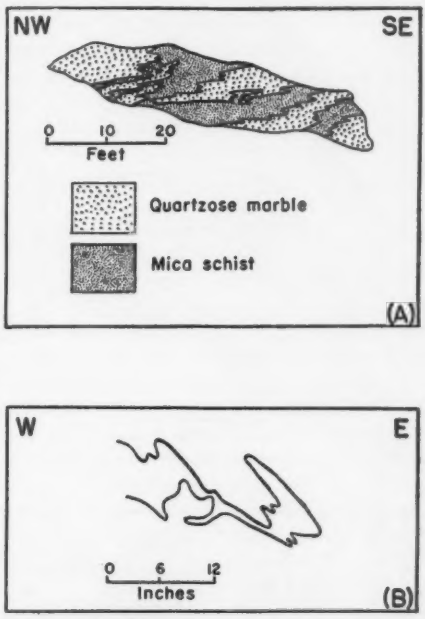


FIGURE 16.—MINOR FOLDS IN THE WAITS RIVER FORMATION

(A) Cross section exposed in a cliff 1500 feet north-northwest of the junction of Pike Hill Brook and Waits River, west of East Corinth. The tapering lens of quartzose marble in the center may be an earlier syncline. The minor crenulations are typical later minor folds. (B) Outcrop just north of road and north of a point at altitude 820 feet in Hedgehog Brook, southeast Topsham. The line represents a single horizon in a thick quartzose marble bed as viewed on the sloping surface of the outcrop.

the average trend of bedding makes a large angle with the axial planes of folds, and both dextral and sinistral patterns are found.

Superposition of later minor structural features on earlier.—Superposition of later minor structural features on earlier is a characteristic feature of the belt that contains slip cleavage, and has various manifestations. Slip cleavage everywhere cuts, and commonly offsets, the earlier schistosity, and the earlier schistosity wraps around the noses of the later minor folds. Such phenomena are fairly common in

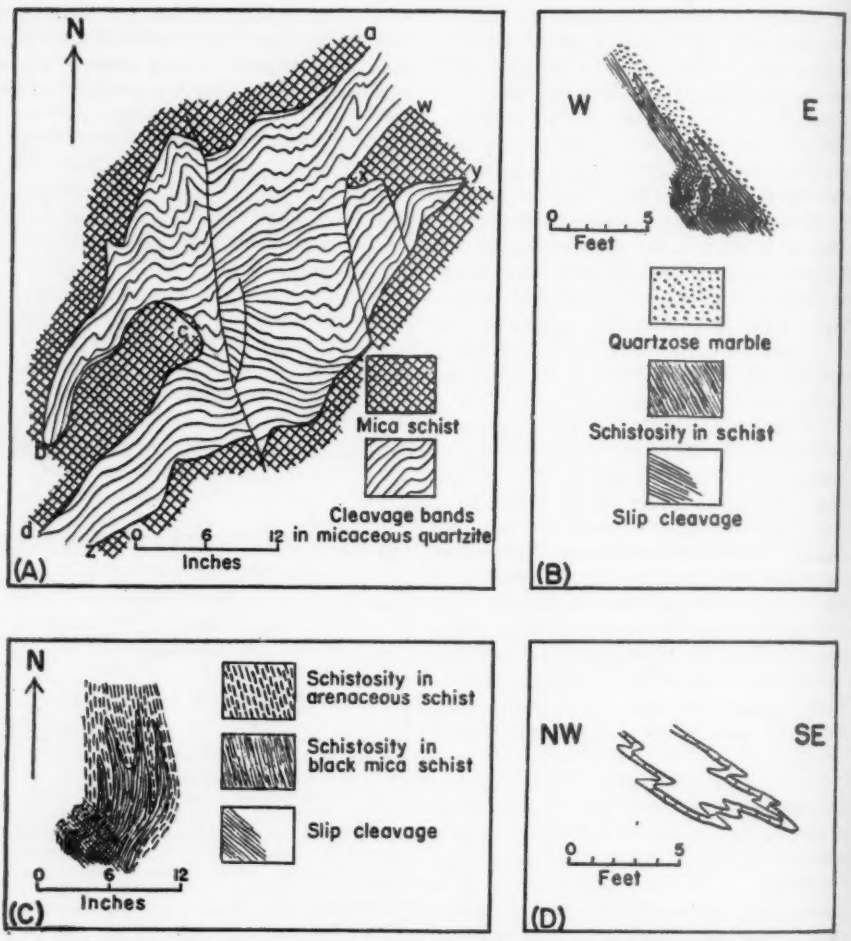


FIGURE 17.—SUPERPOSITION OF LATER MINOR FOLDS ON EARLIER MINOR FOLDS

(A) Micaceous quartzite bed in south-central part of outcrop sketched in Figure 18. Letters refer to points discussed in text. (B) Outcrop in brook just west of boundary between Waits River and Gile Mountain formations, ¼ mile north of the Woodsville quadrangle in the St. Johnsbury quadrangle. (C) Outcrop 4000 feet south-southeast of Red School, Ryegate. (D) Folds in a quartzose marble bed in cliff face. Outcrop 2500 feet northeast of School No. 3, Ryegate.

metamorphic terranes, and may be subject to various interpretations (Mead, 1940, p. 1011, 1020–1021; Cloos, 1937, p. 68–69). Less commonly described are the transection of the axial planes of earlier minor folds by later cleavage and sharp folding of the axial planes of earlier minor folds. (Fig. 17).

In Figure 17A, the line *abcd* represents the western boundary of a bed of micaceous quartzite in mica schist, and line *wxyz* the eastern. The wavy lines are cleavage bands that were originally more or less parallel to the axial planes of earlier-stage folds. The axial plane (or axial surface) of a small minor fold of the earlier stage follows one of the wavy lines connecting *c* and *y*. The axial plane of the com-

plementary fold follows a wavy line connecting b and x. The pattern of this fold-pair is sinistral in plan. Superposed on the northeast-trending axial planes of these earlier folds are later dextral folds whose axial planes strike more or less north. The axial planes of the earlier folds are deformed in the later.

Similarly deformed axial planes are evident in the other sketches of Figure 17. The earlier schistosity of Figure 17B, which dips steeply east to vertical, is parallel to the axial planes of sinistral minor folds of the earlier stage. These folds are revealed by the interfingered marble and schist. The earlier schistosity and the axial planes of the earlier minor folds are both deformed in a later set of dextral minor folds whose axial planes dip gently east, parallel to the slip cleavage. The sketch of Figure 17C is similar except that the earlier minor folds are revealed by the interfingering of mica schist and arenaceous schist, and their pattern is not evident from this small exposure. In Figure 17D, the limbs of an earlier fold with southeastward-dipping axial plane are deformed by small later folds; the axial planes of these later folds have a more gentle apparent dip to the southeast than the axial plane of the earlier fold.

The lime-silicate bands of Figure 18 reveal many earlier minor folds whose axial planes are more or less parallel to the schistosity. A number of such folds with sinsistral pattern are evident in the northeast quarter of the outcrop. In the group of dextral minor folds just south of the center of the outcrop, on the other hand, the schistosity and lime-silicate bands are both folded together. The axial planes of these later folds trend north to northwest, and make a fairly high angle with the schistosity.

The deformed axial planes of the Woodsville quadrangle are significant for two reasons. First, they show clearly that there has been a sequence of tectonic events. One might argue, for example, that the cleavage here called "earlier schistosity" is primarily mimetic after bedding and does not represent a tectonic event. If this were true, of course, the fact that the earlier schistosity wraps around the noses of the later minor folds would have no tectonic significance. The earlier minor folds, however, imply movement as the cause of the earlier schistosity, and, particularly where the axial planes of these earlier folds are deformed, this earlier movement is clearly dated as prior to the later minor folds.

Second, the dominantly sinistral patterns of the earlier minor folds, contrasted with the dominantly dextral patterns of the later, indicate that the events whose order has been established are drastically different. There must have been significant reversal of the direction of differential movement over a fairly large area. This reversal is emphasized and unequivocal where later dextral minor folds are superposed on the axial planes of earlier sinistral folds.

Major folds.—Two major folds, the Wrights Mountain fold and the Ryegate anticline, are revealed by irregularities in the boundary between the Waits River and Gile Mountain formations. Their relationship to the minor structural features described above provides clues to their nature and relative age.

The Wrights Mountain fold is outlined by an abrupt change in the trend of the boundary of the Waits River formation on the southwest slope of Wrights Mountain (Pl. 1). On the northeast limb of this fold, the bedding and earlier schistosity trend north and dip more or less vertically. On the southwest limb, which may be regarded as extending essentially to the western border of the quadrangle, the same structural elements trend N. 70°E. and have an average dip of 15–20°NW. The trace of the axial plane of the fold may be roughly located by drawing a line connecting the points at which the strike of bedding swings from south to west-southwest (Fig. 8, Pl. 3). This line passes just east of Bradford Center, along the east side of the Waits River valley, and strikes about N 30°W. The dip of the axial plane, as revealed in cross sections S-S′ to W-W′, is of the order of 30° to 50°NE. The plan view of this fold is deceptive in that the trace of the axial plane strikes more or less normal to the southwest limb of the fold, and therefore makes this limb look more like a nose than a limb; actually, because of the gentle dip of the axial plane, the solid angle between the plane and each limb is approximately the same, as can be readily determined on a stereographic net. The plunge of the fold is about 20°N.

The axial plane of the Wrights Mountain fold

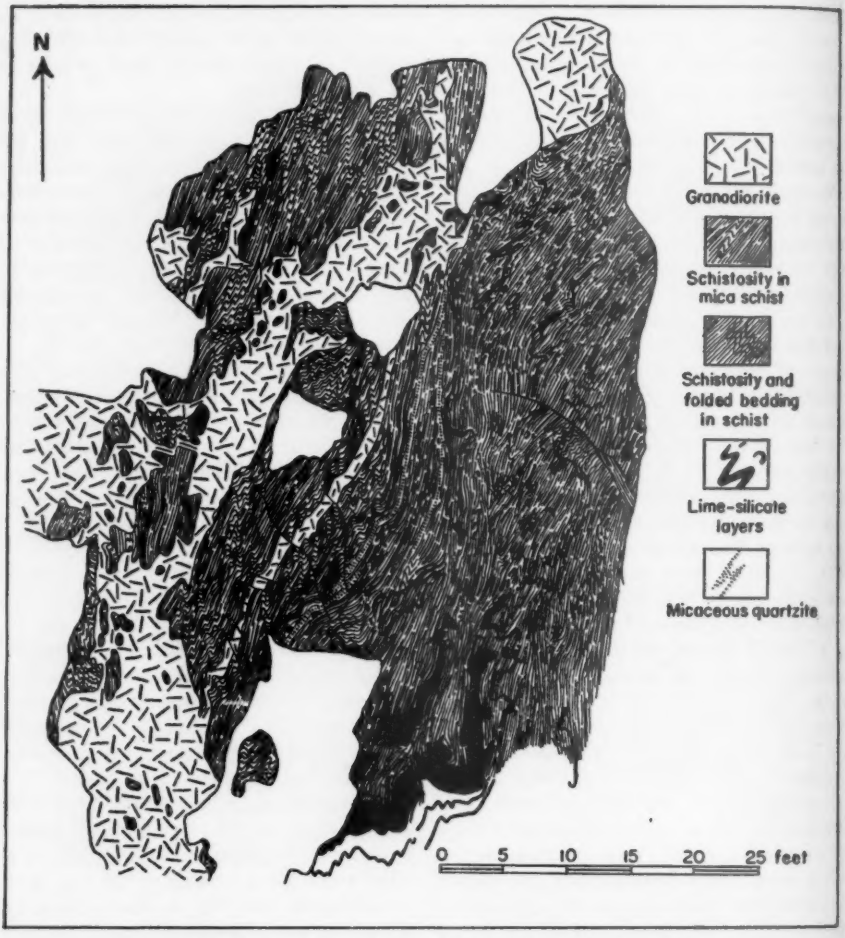


FIGURE 18.—COMPLEX FOLDS OF THE EARLIER STAGE OF DEFORMATION

Later minor folds, superposed on the earlier, have axial planes that trend slightly west of north, and are most abundant in the western half of the outcrop. Bedding and schistosity dip 20°–35°E. Outcrop 800 feet north-northwest of highest point on east-west trail, north side of Blue Mountain.

is more or less parallel, in strike and dip, with the later slip cleavage and the axial planes of the later minor folds (Pl. 3). Its axis is essentially parallel to the plunge of the later minor folds. The earlier schistosity, furthermore, curves with the bedding around the nose of the fold. There can be little question, therefore, that the Wrights Mountain fold was formed during the later stage of deformation.

In the Ryegate anticline, a mile northwest of Ryegate village (Pl. 1), the boundary between the Waits River and Gile Mountain formations is doubled back to form a "V" pointing slightly east of north. The axial plane of this fold strikes about N. 30°E. and is interesected almost at right angles by the axial planes of the later minor folds and slip cleavage (Fig. 8; Pl. 3). The Ryegate anticline is therefore older than

72°15'
44°15'
10'
05'

Ricker Pond
Ricker Mills
Lower Symes Pond
Blue Mtn.
Groton
Ryegate
East Ryegate
South Ryegate
Ticklenaked Pond
Wells River
East Topsham
Newbury Center
Round Pond
Harriman Pond
Newbury
West Newbury
East Corinth
Goulund Hill
CALEDONIA CO.
ORANGE CO.
MONROE FAULT
WELLS THRUST
HOLLOW THRUST
SCOTCH THRUST
AMMONOOSUC THRUST
Dike Complex
River

EXPLANATION

LITHOLOGIC UNITS

- d — Glacial drift
- rg — Ryegate granodiorite
- Oal — Albee formation
- Oo — Orfordville formation
- Om — Meetinghouse slate
- Ogl / Og — Gile Mountain formation, Ogl, calcareous members
- Ow — Waits River formation

SYMBOLS

Strike and dip of planar elements

INCLINED *VERTICAL*

- Bedding
- Schistosity
- Bedding parallel to schistosity
- Slip cleavage
- Axial plane of fold
- Schistosity parallel to axial plane of fold
- Slip cleavage parallel to axial plane of fold
- Schlieren or gneissic banding in granodiorite

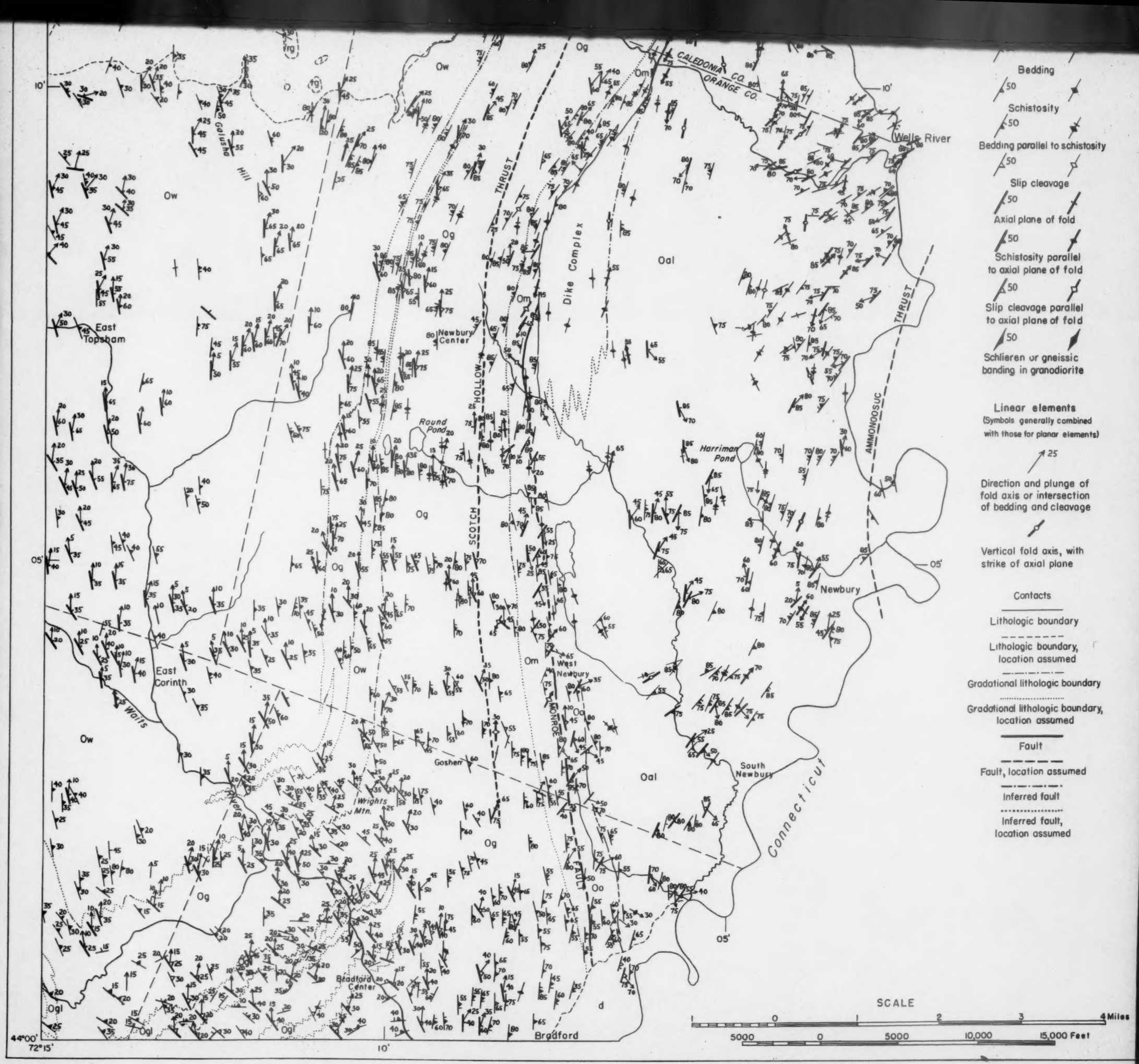
Linear elements

(Symbols generally combined with those for planar elements)

- 25 — Direction and plunge of fold axis or intersection of bedding and cleavage
- Vertical fold axis, with strike of axial plane

Contacts

- Lithologic boundary
- Lithologic boundary, location assumed
- Gradational lithologic boundary



TECTONIC MAP OF VERMONT PORTION OF WOODSVILLE QUADRANGLE, VERMONT-NEW HAMPSHIRE

t
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the slip cleavage and presumably belongs to the earlier stage of deformation.

This major fold is interpreted as an anticline on the basis of its outcrop pattern and the probable northward plunge of its axis. All but one of the earlier minor folds in the quadrangle plunge northward (Fig. 14), and the average plunge is about 32°. An earlier major fold, such as this anticline, should have a similar plunge. Approximately this same plunge may also be derived from the structural data obtained in the immediate vicinity of the fold (Pl. 3). By projecting the boundary between the formations downward parallel to the axes of minor folds, as was done in preparing the cross sections (Pl. 2, C-C′ and D-D′), one can determine the probable shape and attitude of the fold in depth. The basic assumption involved in such projection is almost certainly valid where the major fold is not isoclinal and where the minor folds all have a fairly uniform direction of plunge, as here, because the axes of all minor folds, regardless of their relative age, must lie parallel to the bedding. The Ryegate anticline, as reconstructed in depth by such projection, has a plunge of 29°N. 11°E.

West of the Ryegate anticline, the contact between the Waits River and Gile Mountain formations trends northwest, and the change in strike at a point 1.5 miles due west of Ryegate village suggests the nose of a syncline. Detailed mapping, however, showed clearly that this northwest-trending boundary between formations strikes nearly at right angles to the trend of bedding in the Gile Mountain formation on the west limb of the Ryegate anticline. The boundary appears to mark a fault.

Interpretation of major structure and relative age of formations.—It will be noted that the Wrights Mountain fold and the Ryegate anticline give contradictory evidence of the relative age of the Waits River and Gile Mountain formations. Both folds plunge north and have the Waits River formation in their cores, but the fold at Ryegate is anticlinal in appearance and the Wrights Mountain fold synclinal (Pl. 2). One or the other is obviously unreliable, and, for lack of other unequivocal evidence for relative age, the sequence of formations must be decided on the basis of an interpretation of the relationship of the folds to one another.

The structural relations evident in the western part of the Woodsville quadrangle may have come about in one of at least two ways (Fig. 19). Under the first hypothesis, the Waits River formation is younger than the Gile Mountain, and during the earlier stage of deformation is involved locally in a recumbent fold (Fig. 19, R), the synclinal portion of which is here called the Ryegate anticline. The relations at the end of the earlier stage are shown by the partly solid, partly dashed line AB. During the later stage, the rocks on the east are rotated upward to form the northeast limb of the Wrights Mountain fold (W), producing the present relations represented by the solid line AWB′.

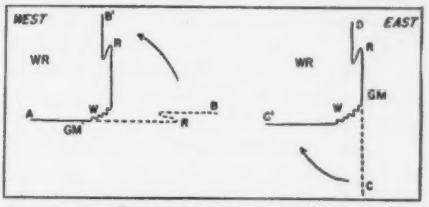

FIGURE 19.—DIAGRAMMATIC CROSS SECTIONS OF STRUCTURE WEST OF MONROE FAULT

Sections illustrate two possible origins, described in text, of present relationship between Ryegate anticline (R) and Wrights Mountain fold (W). Present boundary between Waits River (WR) and Gile Mountain (GM) formations represented by line AWB′ under first hypothesis, and C′WD under second.

Under the second hypothesis, the Waits River formation is older than the Gile Mountain. The rocks are steepened during the earlier stage of deformation, and the Ryegate anticline (R) is formed as part of a very large drag fold on the east limb of an enormous anticline, whose axis lies somewhere to the west. During the later stage, the rocks down the dip from the Ryegate anticline are displaced relatively upward and westward to form the Wrights Mountain fold (W). Relations before and after the later stage of deformation, under the second hypothesis, are shown by lines DC and DWC′, respectively.

The first hypothesis appears offhand to be the simpler and perhaps more reasonable explanation. The strong rotation of the northeast limb of the Wrights Mountain fold offers a

ready explanation for the transection of earlier by later cleavage on that limb, and for the apparent reversal of the direction of differential movement. This first hypothesis is not favored here, however. If intensity of development of the later minor folds and cleavage is proportional to the amount of gross deformation of the rocks, it is most logical to infer that the southwest, rather than the northeast, limb of the Wrights Mountain fold has undergone the greater amount of disturbance. As described above, later cleavage and minor folds are practically nonexistent near the Monroe fault, on the northeast limb, and become more and more intensely developed westward; in the southwest corner of the quadrangle, slip cleavage grades into a transverse schistosity, apparently obliterating the earlier bedding-schistosity. The minor structural features of the later stage, therefore, are best explained by the second hypothesis, and the Waits River formation is regarded as older than the Gile Mountain.

The orogenic movements involved in the second hypothesis are best considered in the light of the geologic structure farther to the southwest and west and will not be taken up here. A more regional treatment is given the subject by White and Jahns (1950).

Scotch Hollow thrust.—The Scotch Hollow thrust strikes nearly north-south and can be traced from Goshen village to Ryegate village. South and north, respectively, of these points its presence is not definitely known. The evidence for its existence is three-fold: (1) Along the line of the fault the bedding, schistosity, and slip cleavage dip more gently westward than elsewhere. This gentler dip suggests drag on a fault. (2) A continuous section in a brook crossing the fault 1¼ miles north of Newbury Center displays about 100 feet of crushed and broken rock. (3) The fault as mapped on structural evidence alone coincides with the eastern limit of garnet-bearing rocks for 3½ miles south of Ryegate (Pl. 1) and closely follows the eastern limit of biotite-bearing rocks from north of Newbury Center to Goshen. Such a relationship is to be expected along post-metamorphism faults (*e.g.*, the Ammonoosuc thrust), which may bring together rocks of different metamorphic grade.

Where the biotite and garnet isograds are closest and the fault is best developed, in Scotch Hollow, the breadth of outcrop of the steeply dipping Gile Mountain formation is least. Such relations indicate crustal shortening, which is accommodated by thrust rather than gravity faulting. The drag indicates that the thrust is relatively from the west (Pl. 2, Section M-M′). When a bed or schistosity plane is rotated by drag along a fault, its attitude approaches that of the fault plane as a limit. The lowest westward dips of bedding in proximity to the Scotch Hollow thrust are about 40° at Newbury Center. The dip of the fault, therefore, is probably slightly less than 40°W. The post-metamorphism Ammonoosuc thrust dips about 36°W., as described above.

The fault cannot be traced north of Ryegate or south of Goshen. If it is assumed that it dies out beyond these points, the breadth of outcrop of the Gile Mountain formation at Goshen represents the breadth before faulting. At this place the formation is about 12,000 feet wide. The breadth of outcrop is narrowest east of Jefferson Hill, where it is about 8200 feet. The horizontal displacement normal to the strike of the fault is therefore of the order of 3800 feet. If the dip of the fault is slightly less than 40°, the dip slip is about 5000 feet east of Jefferson Hill and decreases to the north and south.

Other faults.—Two other faults mapped in the Waits River and Gile Mountain formations occur in the vicinity of Blue Mountain. One forms the contact between the Gile Mountain and Waits River formations northwest from the Ryegate anticline (Pls. 1, 2). The evidence for the fault is found in the southern half of its trace where it truncates the trend of bedding and earlier schistosity in the Gile Mountain formation (Fig. 8). Furthermore, although stratigraphic subdivisions within the Gile Mountain formation are not clearly defined, rough boundaries between more and less quartzose units can be drawn locally; along the southernmost 1½ miles of the fault trace (Pl. 1), the rocks of the Gile Mountain formation adjacent to the fault are unlike at different places.

The dip of the fault and the direction and amount of displacement are not known. It strikes parallel to the later cleavage, and is

assumed, in the cross sections, to have the same dip as this cleavage.

A fault with northeast trend has been drawn from a point 3500 feet southwest of the south end of Lower Symes Pond to the northern boundary of the area east of the pond. On the southeast side of the fault trace the bedding and schistosity strike parallel to the fault and dip steeply or are vertical. On the northwest side of the fault the bedding trends northwest or nearly north. Discordance is suggested, and the fault has been drawn between the areas of different trend. There is no evidence bearing on its dip or the direction and amount of displacement.

Ryegate granodiorite.—The structural relations of the Ryegate granodiorite are best displayed in the vicinity of Blue Mountain (Pl. 1). There the granodiorite was intruded in thick sheets, more or less parallel to the regional structure of the stratified rocks. The map shows that all the sheets join to form a single body on the northwest side of Blue Mountain, but outcrops are few in this section, and possibly there are unexposed schist septa here as well.

The upper contact of the large southwestern sheet is well exposed in a cliff above Gibson quarry and dips 20°NE., parallel to bedding and cleavage in the schists above. Biotite schlieren observed in several quarries in this sheet dip, on the average, about 45°NE. The lower contact is not exposed, but mapping indicates that it has the same strike as the schists. Therefore, the body is believed to dip about 45°NE., parallel to the underlying schists. The divergence of the upper and lower contacts may be due to local irregularity or downward enlargement of the body. Along the south side of the mountain, the lower contact dips more steeply.

The higher intrusive sheets are much less clearly defined. The schist septa are full of large and small dikes and sills of granodiorite and, unlike the lower intrusive sheet, the upper sheets contain large inclusions of schist. The prevailing dips of the schists and observed contacts are gentle northeast, however, and it is believed that the general attitude of all the sheets is similar to that of the lowest one.

Small outlying bodies of granodiorite are subconcordant and tabular like the larger bodies.

Detailed maps of some of the granodiorite-schist contacts bring out several facts very clearly.

(1) Although most visible contacts are essentially parallel to bedding and schistosity, the trend of a contact may be at right angles to the bedding.

(2) There is strong evidence that locally, at least, the granodiorite made way for itself by forcing aside the layers of schist. Figure 20 illustrates the results of such a mechanism.

(3) Most contacts are very sharp, and, even in directions parallel to cleavage, there is little tendency for schist to grade into granodiorite. Feldspathized schist was found only on the hill north of Lower Symes Pond, although injection gneisses were seen farther north in the St. Johnsbury quadrangle.

(4) Inclusions of schist in granodiorite are mostly oriented with their long axes parallel to contacts. The orientation is clearly dimensional, inasmuch as the cleavage within oriented fragments may have random orientation (Fig. 21). The cleavage, however, is normally parallel to the elongation of the fragment.

(5) The granodiorite cuts folds in the schistosity (Figs. 18, 21) and therefore followed some at least, of the folding of the schistosity.

The plutonic bodies in the Groton area are poorly exposed. Most of the area is probably underlain by granodiorite, but schist septa are undoubtedly present. The attitude of the stratified rocks north of the drift-covered area (Fig. 8; Pl. 3) suggests wrapping around a large intrusive, but granodiorite is insufficiently exposed to map as a unit on such meager evidence. All that can be said, therefore, is that there are probably large and small subconcordant intrusives in the area, just as on Blue Mountain.

The granodiorite bodies may have been emplaced in one of three ways: (1) by feldspathization of schists in place, *i.e.*, granitization; (2) by stoping on a large or small scale; or (3) by forceful injection.

The principal argument against granitization, apart from lack of evidence of metasomatism, is the fact that, although the country rock in the Blue Mountain area is argillaceous and

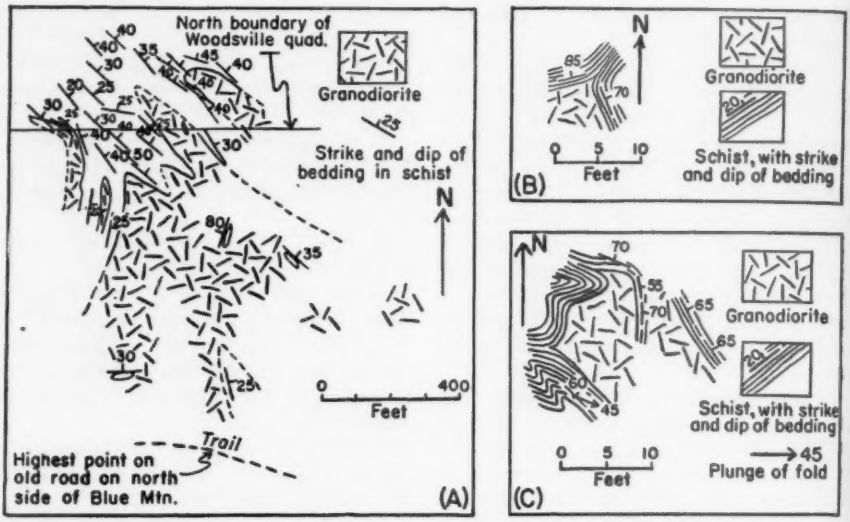

FIGURE 20.—INTRUSIVE CONTACTS OF RYEGATE GRANODIORITE, ILLUSTRATING RESULTS OF FORCEFUL INJECTION

(A) Pace and compass map, north ridge of Blue Mountain. (B) Outcrop beside house 2500 feet south of north border of Woodsville quadrangle on road east of Blue Mountain. (C) Outcrop just south of trail north of Blue Mountain, 500 feet from east end of trail.

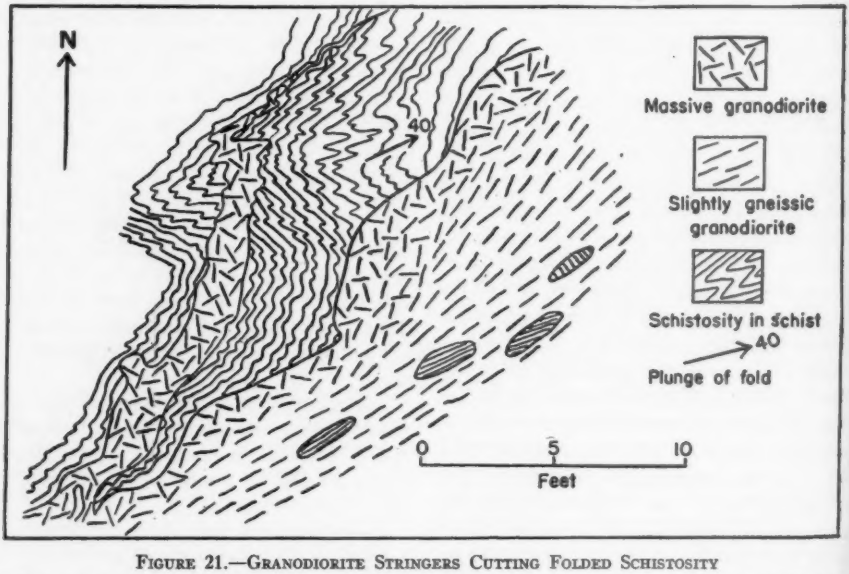

FIGURE 21.—GRANODIORITE STRINGERS CUTTING FOLDED SCHISTOSITY

Inclusions are diagrammatic, but such inclusions are found in the same outcrop outside the field of the sketch. Same outcrop as Figure 20B.

that in the Groton area is calcareous, there is no notable difference in the mineral composition of the plutonic rocks in these two areas.

The bodies lack the characteristic features of those plutons in the region that are believed to have been emplaced by stoping, such as marked truncation of bedding, cleavage, and fold axes (Chapman and Chapman, 1940, p. 201–209).

The remaining mechanism is forceful injection. Evidence that locally the walls have been forced apart is shown in Figure 20. Furthermore, the bodies are lenticular and subconcordant. Schist inclusions are parallel to contacts. Lastly, the trend of the foliation and of the axial planes of later folds diverge around the granodiorite areas (Fig. 8; Pl. 3).

The granodiorite is therefore believed to have been forcefully injected into the schists, mostly along the cleavage planes but partly along fractures of different orientation.

The granodiorite was intruded after most of the folding. Later folds are cut by stringers of igneous rocks (Figs. 18, 21), and no outlying dikes have a secondary foliation, regardless of the angle at which they cut the planes of deformation. A few have a vague foliation parallel to their boundaries, but this is believed to be primary flow structure. And although the granodiorites have unquestionably caused local deformation by their forceful injection (Fig. 20) the dikes in particular are typically cross cutting, with sharp, clean walls, and suggest only that they were intruded into already deformed rocks.

Metamorphism

General Statement

Many of the principal facts concerning the metamorphism have already been mentioned. The present section will therefore merely summarize the data already presented but will also consider some special problems that deserve elaboration. The metamorphism of the mafic dikes is discussed in another paper (Billings and White, 1950).

Progressive Metamorphism

It has been repeatedly emphasized in the description of the metasedimentary rocks that a belt of low-grade metamorphism averaging 4 miles wide extends from the south-central part of the quadrangle to the northeast corner. West and east of this central belt, there is an increase in the grade of metamorphism so that rocks in the western and eastern parts of the quadrangle lie in the staurolite zone. Locally, around bodies of plutonic rocks, the sillimanite zone is attained.

In the western part of the area the porphyroblasts that reflect the progressive metamorphism are later than most of the deformation of the later stage. Large crystals of chlorite, muscovite, biotite, garnet, and staurolite generally cut indiscriminately across slip-cleavage planes, replacing rather than displacing the pre-existing fabric. Microfolds associated with the slip cleavage may be traced uninterruptedly through and across the borders of many porphyroblasts; as seen in thin section, the original outlines of these folds are perfectly preserved by lines of unreplaced opaque inclusions. Where the earlier schistosity is curved in these microfolds, outside of the porphyroblasts, individual flakes of mica that define the schistosity are not themselves bent; large numbers of them have evidently recrystallized or grown as straight plates, oriented parallel to tangents of the fold, after bending had ceased (Richey, 1948, p. 208). Evidence of post-crystallization bending, crushing, or rotation of crystals is sufficiently uncommon and local in the area of slip cleavage to indicate that most of the deformation had ceased before the peak of recrystallization. More substantial overlap of the periods of deformation and metamorphism has been noted west of the Woodsville quadrangle (White and Jahns, 1950, p. 209).

Retrograde Metamorphism

Ammonoosuc thrust.—Rather striking retrograde effects are found directly east of the Ammonoosuc thrust both in the Piermont member and in the main part of tne Albee formation. In a belt 500 to 1000 feet wide on the east side of the fault all the biotite has been converted to chlorite. The retrograde origin of the chlorite is evident because much of it contains rutile needles or leucoxene, both of which are considered to have exsolved during the alteration from biotite. Moreover, as one approaches the

TABLE 9.—ESTIMATED MODES OF ROCKS INVOLVED IN RETROGRADE METAMORPHISM

	1	2	3	4	5	6	7
Number of sections	3	1	2	1	1	3	2
Quartz	44	10	26	45	58	46	3
Orthoclase			20	10		27	
Plagioclase	33	50					45
Sericite			30				2
Chlorite	16	19	19	15	12	11	30
Epidote*	1	12	2	25		t	4
Muscovite	4	1		3	20	10	
Garnet	1				1	2	
Staurolite					7		
Magnetite	x	x	3		2	2	
Ilmenite							1
Pyrite	1			2			
Leucoxene			x			2	2
Apatite	x	x	x	x	x	x	
Tourmaline	t						
Sphene		8					1
Calcite							12
Per cent of anorthite in plagioclase	30	10					10

* Includes clinozoisite.

x—present in 50 per cent or more of thin sections.

t—present in less than 50 per cent of thin sections.

Blank—not seen in any of thin sections.

1. Feldspathic chlorite schist, Albee formation.
2. Feldspathic chlorite-epidote schist, Albee formation.
4. Chlorite-epidote schist, Albee formation.
5. Staurolite-garnet-chlorite schist, Piermont member of Albee formation. Half of staurolite is altered to sericite and chlorite.
6. Feldspathic chlorite schist, Piermont member of Albee formation.
7. Albite-chlorite-calcite rock, retrogressed from amphibolite.

All specimens are from a belt 500 feet wide directly east of the Ammonoosuc thrust.

belt of retrograde metamorphism from the east, biotite crystals in all stages of chloritization may be seen. Staurolite crystals do not begin to alter nor do they disappear as soon as the biotite. Even within 200 feet of the thrust, unaltered cores of staurolite may be found. The altered staurolite has a deep-brown, greasy appearance and forms either a complete pseudomorph or a shell surrounding the amber-colored unaltered staurolite. Thin sections show that the altered material is a fine-grained mat composed chiefly of sericite and what is probably chlorite. Garnet appears to have been more resistant to the retrograde processes than the biotite and staurolite. In specimens where the last two minerals have been completely altered, the garnet may be entirely fresh or chloritized only along a few cracks that cross it. Specimens probably not more than 70 feet from the fault contain relatively fresh garnet. Still closer to the fault, the rocks are cut by irregular quartz veins from 1 to 3 inches wide, and locally a large silicified zone is present.

Amphibolites in a belt several hundred feet wide on the east side of the Ammonoosuc thrust have retrogressed to albite-chlorite-carbonate rocks.

Modes of some of the rocks that have undergone retrograde metamorphism are given in Table 9.

From the descriptions given above, it is apparent that as one goes from the chlorite zone eastward across the Ammonoosuc thrust the first new metamorphic index mineral to appear is garnet, followed by staurolite, and, lastly, biotite. Thus, from such an area one might get the erroneous impression of the sequence in which the index minerals normally appear in progressive metamorphism. It is also important to note that no great amount of shearing appears to have accompanied the retrograde metamorphism. Locally, especially near the thrust, an occasional small fault may be seen in thin sections, but no wholesale shearing has taken place. It is believed, therefore, that the chief instigator of the retrograde metamorphism has been the introduction of aqueous solutions that entered along fractures associated with the faulting.

Hadley (1942, p. 171–173) has described the retrograde metamorphism southeast of the Ammonoosuc thrust in the Mt. Cube quadrangle.

Haverhill-Bradford area.—In Oliverian Brook, 0.8 mile north-northeast of the village of Haverhill, some of the hornfelses contain porphyroblasts of chlorite that are apparently pseudomorphs after biotite. It appears that in this area the biotite zone was attained, but that subsequently the biotite has retrogressed. On a

small knob just west of U. S. Route 5, a mile northeast of Bradford, chlorite pseudomorphs after garnet are present in some of the phyllites. The garnet appears to have formed penecontemporaneously with the widespread earlier schistosity and to have been converted to chlorite during or following the development of the later slip cleavage.

There is thus a suggestion that in the Haverhill-Bradford area there was a stage of metamorphism that antedated the progressive metamorphism in the western part of the quadrangle. The earlier metamorphism in the Haverhill-Bradford area may have been contemporaneous with the progressive metamorphism in the eastern part of the quadrangle. If this is true, the climax of the progressive metamorphism in New Hampshire is older than that in eastern Vermont. A problem for future investigation is to determine the relative ages of the progressive metamorphism in the two states.

Occurrence of Minerals

The mode of occurrence of most of the minerals is essentially the same in the Woodsville quadrangle as in other areas with a comparable degree of metamorphism. A few minerals west of the Ammonoosuc thrust raise special problems that require further discussion.

CHLORITE: A porphyroblastic chlorite was observed locally in the biotite, garnet, and staurolite zones. The indices of four specimens range from $\beta = 1.623$ to $\beta = 1.630$; the optic sign is positive in two and negative in two; birefringence is less than 0.004. According to Winchell (1936, Fig. 4) these chlorites are ripidolite and diabantite, with intermediate FeO/MgO ratio and moderately high alumina content. These chlorite porphyroblasts have the habit of chloritoid and occur as large tabular crystals with ragged edges, lying for the most part at high angles to the schistosity. Where observed, they appear to be the latest crystals to form in the rock. Noble and Harder (1948, p. 958–959) describe porphyroblastic prochlorite from Lead, S. D., and believe that it formed as a product of progressive metamorphism of somewhat greater intensity than that required to form garnets; they show a prochlorite isograd in the middle of the garnet zone. The porphyroblastic chlorites of the Woodsville area though mostly encountered in the garnet and staurolite zones, are also found in rocks of the biotite zone. These chlorites are products of neomineralization, and are unlike the shreddy or ragged chlorites noted below that merely replace biotite and garnet crystals; but where they occur in rocks with garnet, staurolite, and biotite, it is reasonable to assume that they did not form in equilibrium with these minerals. It appears that they are retrograde and were formed during the period of declining temperature that immediately followed the peak of the metamorphism.

Chlorite replaces garnet or biotite in many micaceous rocks and is commonly most abundant as small ragged flakes in and adjacent to planes along which late slipping has taken place. It is abundant in a specimen of breccia from the Scotch Hollow thrust, but elsewhere retrograde chlorite of this type appears to have no systematic distribution or structural significance.

DOLOMITE: The occurrence of dolomite in the Woodsville area is very unusual. In most regions, dolomite reacts with quartz to form tremolite in the biotite, or at most the garnet zone (Harker, 1939, p. 257–258). Qualitative chemical tests of the carbonate of the Waits River formation indicate that dolomite is present with quartz in rocks as far west as the actinolite isograd (Pl. 1), which lies several miles west of the staurolite isograd of the aluminous rocks. The actinolite has undoubtedly originated largely from the reaction of quartz and dolomite in the normal manner, but the temperature at which this reaction took place must have been abnormally high. Thin calcareous beds close to staurolite schist along the eastern boundary of the Waits River formation do not contain actinolite. One is faced with the inevitable conclusion that for some reason the normal low-temperature dissociation of dolomite was prevented. The reason may be very high confining pressure that reduced permeability and prevented the escape of carbon dioxide. Any attempt to reconstruct the folds above the area suggests that the present level of exposure was once very deep.

CLINOZOISITE: The abundance of clinozoisite in metasedimentary rocks with diopside or

biotite and calcic plagioclase, west of the actinolite isograd, is surprising. Barth (1936, p. 823–825) and Billings (1941, p. 875) believe that calcic plagioclase and epidote may exist in equilibrium. In Eskola's amphibolite facies (Barth, Correns, and Eskola, 1939, p. 352), rocks with diopside and calcic-plagioclase do not contain epidote. In the Woodsville area there is no evidence that the clinzoisite is of retrograde origin and out of equilibrium with the other minerals that were formed at the peak of the metamorphism.

DIOPSIDE: The composition of diopside in any specimen of lime-silicate rock is in direct ratio to the composition of actinolite in the same specimen, as determined from refractive indices. Where $\beta = 1.690$ in diopside, it is 1.635 in actinolite, and $\beta = 1.700$ in diopside corresponds to 1.650 in actinolite. Where β of diopside exceeds 1.700, indicating a hedenbergite content of 40 per cent (Winchell, 1933, p. 226), hornblende is present rather than actinolite. Under the conditions of metamorphism in the Woodsville area, therefore, actinolite with a ferro-tremolite content greater than 40 per cent does not form.

Cause of Metamorphism

The problem of the causes of metamorphism is fundamentally the problem of the source of heat. Although mineral equilibria and the attainment of equilibrium are modified by hydrostatic pressure, stress, and the presence of moving solutions, the surfaces whose traces are the biotite, garnet, and staurolite isograds correspond closely to isothermal surfaces, and for practical purposes the deviations caused by factors other than temperature can probably be ignored where the temperature range is large. Three possible sources of heat are considered below.

The rocks of east-central Vermont are most highly metamorphosed near the crestal region of the cleavage arch west of the Woodsville quadrangle (White and Jahns, 1950), and the metamorphism decreases progressively both east and west from this region. The north-south variation along the crestal region has not yet been studied. Large intrusives have been mapped in this crestal region, and their close relation to the principal metamorphism in time and space is striking. If one postulates that these intrusives are apical stocks of a more extensive batholith extending north-south below the cleavage arch, all the features of the regional metamorphism are readily explained.

Billings (1937, p. 557–559) has shown that in western New Hampshire the progressive increase in the grade of metamorphism toward the southeast is spatially related to the appearance of large bodies of the New Hampshire magma series. Moreover, in the eastern part of the Woodsville quadrangle, the high-grade metamorphism in the sillimanite zone of metamorphism is clearly related to the French Pond granite (Pl. 1).

Friction caused by deformation is probably not an important source of heat. The evidence commonly cited to prove that heat has been largely supplied by friction is the coincidence between intensity of deformation and degree of metamorphism in an area. (*See* Ambrose, 1936, p. 278–284.) Such evidence may involve confusion of cause and effect. The same amount of work will produce more distortion in a more plastic body than in a less plastic, other factors being equal. The increase of rock plasticity with temperature is almost axiomatic among geologists (Harker, 1939, p. 182–184; Leith, 1923, p. 322; Longwell, Knopf, and Flint, 1939, p. 370, 395), although not yet fully confirmed by experimental evidence. Griggs (1942, p. 112) has found that dry rocks are not notably affected by temperatures of a few hundred degrees, but notes (p. 114) a marked reduction in the strength of quartz immersed in sodium carbonate solution at elevated temperature. If plasticity does increase with temperature, other independent variables being approximately equal, the relation of apparent intensity of deformation to temperature may well be that of effect to cause, rather than the reverse, as assumed by Ambrose.

In the Vermont part of the Woodsville quadrangle, intensity of deformation is generally greater with higher temperature, as indicated by the isograds. There is, however, one remarkable exception. Westward from the trace of the axial plane of the Wrights Mountain fold, there is a very marked and abrupt increase in the amount of deformation as recorded by the num-

ber of minor folds per unit area. If the metamorphism was caused by heat of friction, there should be a corresponding increase in metamorphism west of this axial plane. Yet the garnet isograd crosses the trace of the axial plane without apparent deflection, and the staurolite isograd, though poorly defined in this vicinity, appears to curve away from, rather than toward parallelism with this line as it is followed south.

In western New Hampshire, Billings (1937, p. 557) could find no evidence that the intensity of deformation was any greater in the high-grade zone of metamorphism than in the low-grade zone.

The writers, therefore, do not consider that heat generated by friction in deformation was an important cause of the metamorphism in the Woodsville quadrangle.

Finally, lateral variations in the degree of metamorphism might result simply from differential uplift of originally flat-lying isothermal surfaces. There is close correlation in central Vermont (White and Jahns, 1950) between a broad area of domical uplift and a high grade of metamorphism. Inasmuch as this domical uplift is at least locally occupied by intrusive granitic rocks, however, the importance of uplift alone is not readily evaluated. Billings (1937, p. 557) finds no correlation in New Hampshire between domical uplifts and metamorphism.

Heat introduced by rising bodies of magma seems to be the most likely source of heat for the metamorphism.

Tectonic Evolution

The major folds and the earlier minor folds were the first product of the tectonic forces. In general, the Woodsville quadrangle is located on the east flank of a complex anticlinorium. Four synclines are present on the flanks of the anticlinorium (Fig 2): the Monroe (southeast corner of the St. Johnsbury quadrangle), Walker Mountain (directly west of the Ammonoosuc thrust), Salmon Hole Brook (directly west of Northey Hill thrust), and Garnet Hill (directly southeast of Northey Hill thrust).

It has been shown that the earlier minor folds evolved during the later stages of the major folding. The later minor folds were considerably younger.

It has long been realized that the major faults were not contemporaneous with one another (Billings, 1937, p. 525–531). The relations in the Littleton-Moosilauke area indicate clearly that the Ammonoosuc thrust is younger than the folding (Billings, 1937, p. 529). Moreover, it is younger than the progressive metamorphism of the eastern part of the Woodsville quadrangle, inasmuch as some of the metamorphic zones have been eliminated along the fault. It has already been shown that the Scotch Hollow thrust is in many ways similar to the Ammonoosuc thrust and thus presumably contemporaneous with it. The Scotch Hollow thrust is younger than the progressive metamorphism in the western part of the quadrangle. Consequently these two faults are considered relatively young tectonic features.

The Northey Hill thrust is older than the Ammonoosuc thrust, for it is older than the progressive metamorphism of the eastern part of the area (Billings, 1937, p. 531). On the other hand, the major folds were well formed before the Northey Hill fault developed, because in the northern part of the Moosilauke quadrangle the fault cuts across the west limb of the Garnet Hill syncline. It is likely that the fault at one time had a more gentle dip to the west and that additional compression rotated it to its present attitude.

The Monroe fault is believed to have formed during the earlier stage of deformation, but movement must have continued beyond the close of the later stage. Evidence bearing on this subject is as follows:

(1) The fault is parallel, in strike at least, to the earlier schistosity of the rocks on either side; it makes a large angle with the later cleavage and axial planes of the later minor folds.

(2) On the west side of the fault, particularly in the southern part of the area, the schistosity is folded. The axial planes of the resulting minor folds trend northeast and dip steeply southeast. Many of these folds have a slip-cleavage parallel to their axial planes. They are believed to be the same as the similarly oriented later folds with slip cleavage in the Orfordville formation across the fault to the east. The prob-

able magnitude of the fault makes it unlikely that folds and cleavage of identical type and orientation have been brought together from widely separated points, and most of the movement on the fault is therefore believed to antedate these later structural features.

(3) No breccia or broken cleavage planes were found along the fault in the Woodsville area, although Eric (1942, p. 65–66) found these phenomena locally along the fault in the Vermont portion of the Littleton quadrangle. Such features indicating late movement are certainly not abundant.

(4) The biotite isograd is displaced along the fault at Boltonville. Although this apparent displacement might be explained as the result of processes other than faulting—such as the greater metamorphism of the rocks on the east side of the fault by heat from the numerous late-tectonic mafic dikes—it is tentatively regarded here as a real mechanical displacement. The horizontal offset is about 3 miles; if the surface whose trace is the isograd dips eastward at a gentle to moderately steep angle (a logical assumption; *see* Noble and Harder, 1948, p. 969), its intersection with the fault plunges gently north, and the vertical displacement on the fault is only a few thousand feet. The east side is upthrown. A displacement of this order of magnitude is clearly inadequate to explain the stratigraphic break, and the amount of post-metamorphism movement must therefore represent but a fraction of the total.

In summary, the Monroe fault is believed to have formed, and most of the displacement on it to have taken place, during the earlier stage of deformation. There was, however, some post-deformation and post-metamorphism movement on it.

The age of the cleavage has been discussed in detail already. Suffice to say that the schistosity east of the Monroe fault is believed to be contemporaneous with the earlier folding, whereas the slip cleavage in the central part of the quadrangle and the schistosity in the western part are believed to be contemporaneous with the later folding.

The chronological position of the plutonic and hypabyssal rocks can be established only approximately. The metamorphosed mafic dikes are older than the metamorphism. The Bethlehem gneiss is syntectonic as it is strongly deformed. The Haverhill granodiorite, the French Pond granite, and the Ryegate granodiorite are late tectonic and are slightly deformed. The camptonite belongs to the White Mountain magma series and is post-tectonic.

The age of the metamorphism has been discussed on previous pages. Clearly, the final stages of the progressive metamorphism in the western part of the area are younger than the slip cleavage. Possibly, the progressive metamorphism in the eastern part of the quadrangle is somewhat older, but conclusive evidence is lacking.

Geologic History

The recorded geological history in the Woodsville quadrangle began in the Middle Ordovician. The Waits River, Gile Mountain, Meetinghouse, Orfordville, Albee, Ammonoosuc, and Partridge formations were deposited prior to the Silurian. Although the thickness of these formations cannot be stated with precision, it is certain that the total accumulation was many thousands of feet, probably at least 20,000 feet.

These rocks were deformed during the Taconic revolution near the end of the Ordovician, but in this area the intensity of this deformation does not appear to have been great. The proof of such a revolution in central New England is shown most convincingly around Littleton, New Hampshire, where Silurian rocks rest unconformably on the pre-Silurian (Billings, 1937, p. 468). Upper Silurian appears to have been a time of non-deposition, but fossils are scarce and rocks of this age may be present. During extensive sedimentation in the Lower Devonian at least 5000 feet of shales and sandstones accumulated (Billings, 1937, p. 468).

The great orogeny and the associated igneous intrusives are Acadian—that is, middle or late Devonian. The most convincing evidence is found in the Franconia quadrangle (Williams and Billings, 1938, p. 1025–1026), where the unmetamorphosed Moat volcanics of the White Mountain magma series rest unconformably on the highly metamorphosed Lower Devonian Littleton formation. The orogeny must be post-Littleton and pre-Moat. The White Mountain

magma series is comagmatic with the Quincy granite near Boston, Massachusetts; the Quincy is pre-Pennsylvanian. Thus between lower Devonian and Pennsylvanian the following events occurred in central New England (Williams and Billings, 1938, p. 1040): (1) orogeny, including intrusion of the New Hampshire magma series; (2) erosion; (3) extrusion and intrusion of the White Mountain magma series. The first event is assigned to the middle and late Devonian (?), the second to the late Devonian (?) and early Mississippian (?), and the third to the Mississippian (?).

All the complex tectonic events in the Woodsville quadrangle are considered Acadian. None are believed to be Appalachian (Permian) because the Mississippian (?) White Mountain magma series throughout Vermont and New Hampshire has not been subjected to compressional forces (Billings, 1945, p. 44–53). It is believed that none of the structures described in this paper are Taconic (late Ordovician), because the pre-Silurian and post-Silurian rocks of central New England show the same intensity of deformation.

The youngest bedrock in the area is camptonite, a late member of the Mississippian (?) White Mountain magma series.

The geological history between the Mississippian (?) and Pleistocene is shrouded in mystery. Certainly many thousands of feet of rock were eroded. Pleistocene and Recent events are recorded by glacial, fluvio-glacial, and fluviatile deposits, as well as by features resulting from glacial and stream erosion.

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MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, JUNE 20, 1949.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 697-699 JUNE 1951

Short Notes

STRUCTURE OF THE SUGARLOAF MOUNTAIN AREA, MARYLAND, AS A KEY TO PIEDMONT STRATIGRAPHY

By ANNA J. AND GEORGE W. STOSE

A recent paper by D. M. Scotford (1951) deals with an area that we have surveyed and on which we have published (Jonas and Stose, 1938; Stose and Stose, 1946, p. 65–74, 80–81). Scotford's interpretation differs from ours both as to stratigraphy and structure. He regards all the beds of the Sugarloaf Mountain quartzite as Weverton quartzite, as we suggested they might be. The encircling phyllites (the Urbana and Ijamsville) he correlates with the Harpers phyllite, of Lower Cambrian age, which overlies Weverton quartzite in the Blue Ridge uplift west of Frederick Valley. Also he regards the structure of the Sugarloaf Mountain area as anticlinal, not synclinal as we interpret it.

The section at Sugarloaf Mountain as interpreted by the Stoses is as follows:

		Feet
Sugarloaf quartzite Cambrian?	Thick-bedded white quartzite, composed of tightly packed round glassy quartz grains; makes the upper cliffs of Sugarloaf Mountain.	400±
	Poorly exposed, softer quartzite, and banded slate	70±
	Hard, thin-bedded white quartzite, composed of tightly packed round dark glassy quartz grains in part stained by iron oxide and containing thin seams of hematite; makes lower ledges	50±
	Concealed	50±
Urbana phyllite	White quartzite, in part stained bright red, composed of densely packed round quartz grains, and thick-bedded crumbly granular quartzite interbedded with thin layers of slate	50±

Scotford's map (Pl. 1 of his paper) is in large part an outcrop map and is not a geologic map because he has not recognized the lithology of the individual beds of quartzite and has not connected equivalent beds of similar lithology. For instance, the upper cliff-making quartzite is much thicker (400 feet ±) than the others. The next lower quartzite has a similar lithology but is not so thick (about 50 feet) and commonly is stained red from contained hematite. The next underlying quartzite is interbedded with slate and probably should be included as part of the Urbana phyllite. Still lower greenish quartzites are less well assorted and contain angular quartz grains. Had Scotford recognized beds that are similar and connected them, thus making a geologic map, the synclinal structure of the mountain would have been evident even though the beds are in large part overturned.

The quartzite of Sugarloaf Peak as viewed from the road at the south (Fig. 1) appears as the uppermost bed of a syncline, and the dips of the quartzite forming the crest of the Peak support this conclusion. Scotford (p. 64) does not contest this obvious synclinal structure, but in his only structure section (his Fig. 10) he shows no syncline at this place and does not explain its relation to his theoretical anticlinal structure.

The thick beds of quartzite (400 feet ±) that form the Peak are in the center of the syncline and do not extend north or south. We have observed quartzite beds on the slope both west and north of the Peak which dip beneath the massive quartzite of the crest (Stose and Stose, 1946, Fig. 16, p. 73) and have traced these beds almost continuously to the south and east, to connect with similar quartzite on the east side of the mountain. These relations indicate that the structure of all the Sugarloaf Mountain quartzites in the vicinity of the Peak is synclinal.

We have shown an anticline (1938; Fig. 2 of this discussion) which follows Furnace Creek and extends north of the divide into Bear Branch. This anticline corresponds approximately to Scotford's anticlinal axis. The sericitic quartzite and phyllites exposed in this anticline

in Bear Branch and southwestward are lithologically like the Urbana phyllite (Harpers of Scotford), as we have so shown them on our map (1938). We traced them in almost continuous outcrops southward around the synclinal Peak into the Urbana phyllite east of Sugarloaf Mountain. Thus, the Sugarloaf Mountain quartzite is not in the center of Scotford's major anticline.

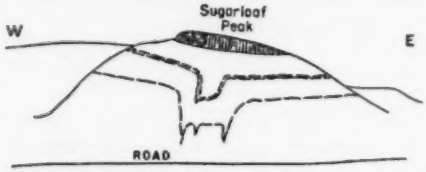

FIGURE 1. SKETCH OF SUGARLOAF PEAK AS SEEN FROM ROAD AT SOUTH, NEAR MT. EPHRAIM
The synclinal character of the top bed is evident.

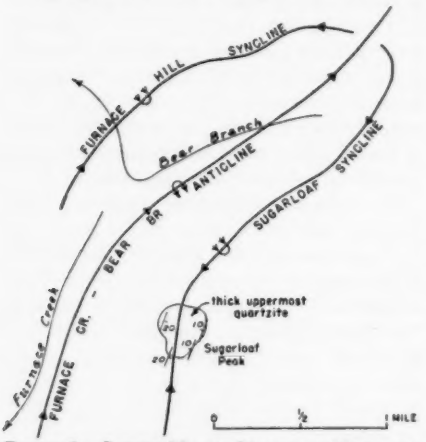

FIGURE 2.—SKETCH MAP OF SUGARLOAF MOUNTAIN
Showing the anticline separating Sugarloaf syncline and Furnace Hill syncline.

We all agree that the beds at Sugarloaf Peak are the uppermost beds of the Sugarloaf Mountain quartzite (Scotford's Fig. 10). If the major structure were anticlinal, as Scotford contends, these massive thick beds should descend on the west side of the anticline into Furnace Hill. They are not present there except at the north crest of the hill. Scotford explains their absence (p. 64) as due to shearing out and thinning of beds on the steep northwest limb of his major anticline. We regard the quartzite in Furnace Hill as largely lower beds exposed in a close syncline (Fig. 2) and separated from the main part of the Sugarloaf Mountain quartzite to the east by the Urbana phyllite brought up on the Furnace Creek–Bear Branch anticline.

Scotford's main argument for the anticlinal structure of the Sugarloaf Mountain quartzite is based on his correlation of the phyllites around the quartzite as Harpers phyllite, and, if the phyllites are Harpers, the Sugarloaf Mountain quartzite must be brought up on an anticline. He correlates the Urbana and Ijamsville phyllites with the Harpers on lithologic similarity, explaining the differences in mineral composition between the phyllites east of Frederick Valley and that of the Harpers in the type locality west of the valley as the result of increased metamorphism eastward. Some of the phyllites near Sugarloaf Mountain do have a composition similar to true Harpers, but most of them contain an abundance of chlorite, and the green phyllites east of the mountain contain also epidote, clinozoisite, and actinolite, minerals not found in the true Harpers. We mapped (1938) these green epidotic phyllites as metabasalt because of their mineral composition and because they extend northward into a series of amygdular basalts and interbedded tuffs of undoubted volcanic origin. These volcanic rocks are interfolded with the Ijamsville phyllite which contains in many places rhyolite flows and smeared out vesicles. Due to shearing, volcanic textures are largely destroyed in the epidotic phyllites near Sugarloaf Mountain, and Scotford, not finding such textures, doubts their volcanic origin. If he had extended his field studies a few miles to the north and east, the origin of the epidotic phyllites and the blue and purple phyllites of the Ijamsville should have been evident to him.

We cannot claim that we discovered the volcanic origin of these rocks. It was recognized in the 1890's by early workers in the Piedmont of Maryland and was noticed by A. Keith when he took over the mapping of that area in 1894, after the death of George H. Williams. The conclusion of their volcanic origin was later confirmed and extended by Mathews (1904) and Mathews and Grasty (1910). The Harpers phyllite throughout its extent contains no effusive rocks, hence has no epidotic phyllites, and

the difference of its mineral composition from that of the phyllites near Sugarloaf Mountain and to the east is due to a difference in origin and chemical composition rather than to a lower rank of metamorphism in the true Harpers. The correlation of the Urbana and Ijamsville phyllites with Harpers phyllite therefore is not valid on lithologic grounds; hence the Lower Cambrian age of the phyllites and the schists of the eastern Piedmont is not proved, as Scotford claims.

The Urbana and Ijamsville phyllites and associated rocks are thrust westward over fossil-bearing Cambrian and Ordovician rocks of Frederick Valley (Jonas and Stose, 1938; Stose and Stose, 1946, p. 126–128). Scotford (p. 63) recognizes this fault from the stratigraphic and structural discordance on the two sides, but minimizes the probable amount of displacement. We regard this fault (Stose and Stose, 1946, p. 126–328) as the Martic overthrust. Cloos and Hietanen (1941) studied the fault near Mine Ridge in Pennsylvania where the fault was named (Knopf and Jonas, 1949, p. 79–80). In that area, folding and metamorphism common to both the overthrust schists and to the overridden lower Paleozoic rocks have obscured evidence of thrusting, and Cloos and Hietanen have thrown doubt on the existence of the thrust. The schists of the Martic block in Pennsylvania (Stose and Stose, 1944, p. 68–72) overlie Harpers phyllite, Antietam quartzite, and Vintage dolomite of Lower Cambrian age, and Conestoga limestone of probable Upper Cambrian and lower Ordovician age. The contact is either an overlap or an overthrust. We believe it is an overthrust.

We have suggested (Stose and Stose, 1950) that the schists, phyllites, and volcanic rocks of the Glenarm series in the Martic overthrust block are the metamorphic equivalents of the Taconic sequence of New England. The Glenarm series is derived from a thick succession of shales and sandstones whose deposition was marked by extensive vulcanism, characteristic of deposits in an eugeosyncline. The Glenarm series in Pennsylvania and in Maryland, east of Frederick Valley, overlie a lower Paleozoic carbonate sequence whose deposition was not accompanied by vulcanism. The carbonate sequence and underlying quartzose sediments, including the Weverton quartzite and Harpers phyllite, were laid down in the Appalachian miogeosynclinal trough which lay west of the eugeosyncline. The Martic overthrust in the Taconic orogeny carried the rocks of the eugeosynclinal trough westward over those deposited in the miogeosyncline. If so, the amount of displacement on the fault on the east side of Frederick Valley was considerable.

To reiterate our conclusions which differ from Scotford's: Sugarloaf Mountain quartzite is synclinal and not anticlinal; the phyllites that surround the quartzite are not Harpers but are phyllites of an eastern shale sequence containing volcanic flows and tuffs, which we have called the Urbana and Ijamsville.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 62, PP. 700-701 JUNE 1951

MECHANISM OF WEATHERING

BY DUNCAN MCCONNELL

Under the title *Mechanism of weathering*, A. F. Frederickson (1951) has presented a novel explanation of the interaction between water and albite. Regrettably he has called upon several poorly understood phenomena and their more complex interrelations to explain a phenomenon which may be simpler than any of the several processes which he has combined in his explanation. For example, in outlining a method for removal of Na ions by aqueous solution from albite, one is asked to accept (1) a quartz structure for the Helmholtz double layer adjacent to the feldspar surface, (2) an adequate fit between the feldspar and ice structures, (3) migration of hydrogen ions into the lattice of feldspar and into regions already occupied by sodium ions having repulsive charges, (4) expulsion of sodium ions through the feldspar structure into and through the ice layer, etc. Insofar as this hypothesis may contain elements of realism, it deserves consideration. Nevertheless, some violence was done to existing terminology, and certain portions of the discussion contain incompatibilities which were not adequately treated. Further consideration of the subject therefore seems desirable.

Although numerous theoretical concepts are discussed, such as the results of Bernal and Fowler (1933), little use is made of these hypotheses in Frederickson's explanation of the weathering process. Actually, Frederickson's hypothesis is extremely simple: Protons migrate into the structure of albite and, having gained entrance to the vicinities of Na ions, repulse the Na ions, which leave the albite to enter the aqueous phase. The motivating force, which induces protons to enter the albite structure, is not clearly indicated. Furthermore, he does not discuss and eliminate an alternative proposition: That the Na ions escape first and are instantaneously replaced by H ions from the water. The significantly larger radius of the K ion immeasurably increases the difficulty of handling potash feldspars by identical methods, if one assumes, as Frederickson apparently does, that these ionic migrations can occur at depths of more than a few unit cells within the crystal.

One notes (p. 224): "On the basis of the X-ray diffraction pattern of water, it has been deduced that a certain regularity of arrangement of the molecules in liquids exists—much more than in gases and slightly more than in some solids."

This statement provides ample opportunity for misinterpretation if one presumes that, here as elsewhere, he refers to crystalline solids rather than glass or some other poorly organized solid which is energetically unstable.

One obtains the impression (p. 230) that some fundamental relationship exists between the $AlSi_3O_8$ trellis framework of the feldspar and the Si_2O_5 sheet structure of clays, and that transformation of the former structure into the latter can be accomplished merely by removal of Na ions from albite. Although Footnote 7 indicates that the structures are different, it fails to indicate the major changes required by this process. Surely a complete structural reorganization of this type, involving as it does a different co-ordination number for some of the aluminum as well as rearrangement of the SiO_4 tetrahedra, seems more difficult to visualize than does the mere probability of escape of Na ions from the surface of albite crystals by methods less complex than base exchange between two crystalline phases.

Far more serious is his use of the term "base exchange" in describing the dissolution of feldspar in water. This erroneous concept is not new. Wiegner (1931) set down the following expression:

$$KAlSi_3O_8 + H_2O \rightarrow HAlSi_3O_8 + KOH$$

which has been widely quoted in literature on soils. This expression completely lacks realism because the existence of $HAlSi_3O_8$ as a crystalline phase has never been demonstrated. Under any circumstances base exchange is a reversible process, and even Wiegner presumably does not conceive of the hydrolysis and re-formation of

feldspar as a reversible reaction which takes place instantaneously at ordinary temperatures in aqueous media.

Walton (1949) and Calmon (1949) have supplied numerous details on cation-exchange (*i.e.*, base-exchange) processes which need no reiteration here except to illustrate the principle of reversibility mentioned above. A simple example obtains in the use of zeolites for water treatment:

$$Na_2Zeolite + Ca^{++} \rightarrow CaZeolite + 2Na^{+} \text{ (softening)}$$

$$CaZeolite + 2Na^{+} \rightarrow Na_2Zeolite + Ca^{++} \text{ (regeneration)}$$

The important consideration involved is the fact that this softening-regeneration cycle can be repeated many times without serious deterioration of the solid phase and can be brought about more or less at the whim of the experimenter through control of the ionic concentrations (actually, ionic activities in thermodynamic calculations) of the aqueous phase. These principles surely do not obtain in the case of interaction between feldspar and water, and no useful purpose is served by extending the use of "base exchange" to cover this and all other types of chemical reactions in which an alkali or alkali earth moves from one side of a chemical expression to the other, accompanied by the opposite movement of another similar ion. The confusion between cation exchange and adsorption, for example, has been pointed out by Ross (1943).

Also difficult to comprehend is Frederickson's summary statement: "The mechanism of weathering is essentially a base-exchange process between two similar crystal structures with different degrees of order." Without attempting to question by what extension of the meaning of "base exchange" such a process can take place between two crystalline phases, one wonders why the entire discussion of the breakdown of the feldspar crystal has been restricted to consideration of the Na ion. Dissolution of Al and Si takes place simultaneously although at slower rates. The quite erroneous impression is created with Al and Si remain entirely associated in a solid phase, which is capable of autogenous structural rearrangement after the Na ions have left. This conclusion is quite at odds with the experimental data obtained by Armstrong (1940) and other investigators, but no explanation of these marked discrepancies was furnished. Not only did Armstrong find that significant amounts of Al and Si were released during the dissolution of feldspar in water, but the fact that these constituents passed to the cathode chamber during electrodialysis suggest their presence as ionic constituents, in part at least.

Although quite original in the manner of integrating the numerous hypotheses discussed, Frederickson's explanation leaves much to be desired by way of demonstrating through observational data that these several concepts can be linked together in any such manner. His use of "base exchange" in describing the dissolution and hydrolysis of feldspar is regrettable. "Base exchange" (now usually called cation exchange) has had a useful application in industrial chemistry for many years, and it has almost universally been applied to a reversible interaction between a solid and a liquid. Admittedly Wiegner and a few others have contributed greatly to the confusion which now exists.

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